

UNION GÉODÉSIQUE ET GÉOPHYSIQUE INTERNATIONALE
INTERNATIONAL UNION OF GEODESY AND GEOPHYSICS

ASSOCIATION INTERNATIONALE
D'HYDROLOGIE SCIENTIFIQUE

INTERNATIONAL ASSOCIATION
OF SCIENTIFIC HYDROLOGY

ASSEMBLÉE GÉNÉRALE DE HELSINKI

25-7 — 6-8 1960

GENERAL ASSEMBLY OF HELSINKI

COMMISSION DES NEIGES ET GLACES
SNOW AND ICE COMMISSION

PUBLIÉ AVEC L'AIDE FINANCIÈRE DE L'UNESCO

PRIX : 350 Frs belges

PUBLICATION N° 54

DE L'ASSOCIATION INTERNATIONALE D'HYDROLOGIE SCIENTIFIQUE

SECRÉTAIRE : L.J. TISON
BRAAMSTRAAT 61, (RUE DES RONCES)
GENTBRUGGE (BELGIQUE)
1961



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INTERNATIONAL ASSOCIATION
OF SCIENTIFIC RESEARCHERS
IN THE FIELD OF
ASTRONOMY AND
COSMOLOGY

ASSEMBLÉE GÉNÉRALE DE HELSINKI

1954

GENERAL ASSEMBLY OF HELSINKI

COMMISSION DES SCIENCES ET DE LA
COSMOLOGIE

REPORT ON THE WORK OF THE COMMISSION

1954

1954

IN THE
CITY OF HELSINKI
FINLAND
1954

COMMISSION OF SNOW AND ICE

COMMISSION DE NEIGE ET GLACE

RESOLUTIONS MADE AT THE HELSINKI ASSEMBLY 1960

I. AT A SPECIAL MEETING HELD AT THE REQUEST OF THE PRESIDENT OF THE ASSOCIATION OF SCIENTIFIC HYDROLOGY

<i>Present; —</i>	Dr. H. G. Wilm	President of the Association
	Prof. L. J. Tison	Secretary of the Association
	Prof. R. Finsterwalder	President of the Commission
	Mr. W. H. Ward	Acting Secretary of the Commission
	Prof. G. Avsiuk	U. S. S. R.
	Prof. P. A. Shumsky	U. S. S. R.

The meeting was held to discuss a Russian proposal to form a separate association for glaciology and to expand the international organisation of glaciology.

It was agreed that: —

- a) No Association for glaciology shall be formed in the next 3 years.
- b) The President of the Commission of Snow and Ice shall be a member, ex-officio, of the Council of the Association.
- c) The Commission shall hold a symposium on the variation of the regime of existing glaciers in 1962.
- d) The National Committees for Scientific Hydrology shall be asked to appoint sub-committees for snow & ice, and National correspondents to the Commission of Snow & Ice.

II. AT THE TRIANNUAL BUSINESS MEETING OF THE COMMISSION OF SNOW & ICE

It was agreed that: —

- a) The Commission shall undertake the permanent task of recording the variations of existing glaciers, the following *sub-committee on Variations of Existing glaciers* being elected: —

A. Bauer, (Chairman) France	P. Kasser, Switzerland
G. Avsiuk, U. S. S. R.	P. Meier, U. S. A.
T.J. Blachut, Canada	J. F. Nye, U. K.
R. Finsterwalder, Germany	Solaini, Italy.

The terms of reference of the sub-committee shall be: —

"The prepare a document detailing the measurements to be made on existing glaciers in all countries so as to record their variations from time to time, to recommend the various means by which this work may be accomplished, and the results collected together and published. This document to be submitted to the Secretary of the Commission by 1st. February 1961".

b) The International Committee on Geophysics may wish to consider using the existing World Data Centres (Glaciology) for assembling the data arising from item a) above when their present task with I.G.Y. is complete.

c) A symposium entitled 'The Variations of the Regime of Existing Glaciers' shall be held in Obergurgl, Austria, in September 1962. Prof. H. Hoinkes (Innsbruck) agreed to act as local organiser.

d) The National Committees of Scientific Hydrology shall be asked to appoint National Correspondents to the Commission.

e) The following four divisions of the Commission shall be formed for the purpose of developing their own programmes of work and for organising meetings: —

1. Division on Glaciers
2. » » Seasonal Snow Cover & Avalanches
3. » » Sea, Lake & River Ice
4. » » Ground Ice.

The existing officers shall be responsible directly for Division 1; M. de Quervain to be responsible for Division 2; E.R. Pounder for Division 3; and M. J.A. Bender for Division 4 for the next 3 years.

f) The following persons shall hold office for the next 3 years: —

President. P. A. Shumsky (U.S.S.R.)

Vice-Presidents. A. Bauer (France), W.O. Field (U.S.A.), G. Morandini (Italy)

Secretary. W.H. Ward (U.K.)

GLACE DE MERS ET DE LACS — NEIGE
SEA AND LAKE ICE — SNOW

CRYSTAL STRUCTURE OF BRACKISH AND FRESH-WATER ICE

PALOSUO, ERKKI

Institute of Marine Research, Helsinki, Finland

SUMMARY

In the Baltic Sea the ice in the estuaries of the rivers is fresh water ice, with the optical *c*-axes of the ice crystals oriented vertically. Farther out in the area of brackish water the *c*-axes are horizontal. In the northernmost part of the Gulf of Bothnia, where the salinities are extremely low, the areas covered by these two types of ice clearly are separated by a line of demarcation.

In the later part of the paper, tests are described on the influence of various factors on the orientation of the crystal axes in freezing.

1. ICE IN THE BRACKISH WATER

It is generally known^(1,2,3) that in pure water ice the crystals are oriented with their *c*-axes vertical, but in the sea ice the *c*-axes are horizontal. This concerns mainly sheet ice⁽⁴⁾, which has been developed from ice skim by thickening. In agglometric ice consolidated from slush or pan, the structure is irregular.

In the Baltic the salinity gradually decreases northward. In the northernmost part of the Gulf of Bothnia the salinities are extremely low and in the estuaries of the rivers the water is fresh. This is the reason why the area off Tornio (Fig. 1) was chosen for the investigation of ice crystal orientation.

Ice samples were taken at Röyttä, where the water is almost fresh. A few kilometers farther southward the next sample was taken, and so on for about ten kilometers.

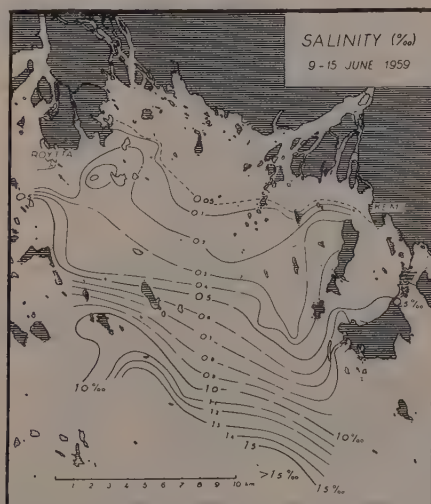


Fig. — The surface salinities of the water off Tornio in the summer of 1959 observed on board of R/V Aranda. (Unpublished report by Ilmo HELA, Institute of Marine Research, Helsinki).

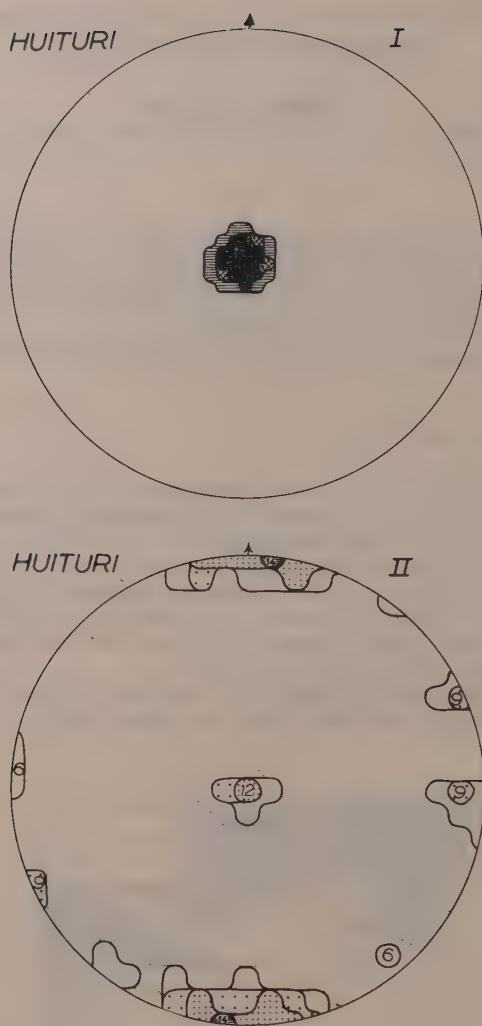


Fig. 2-3 — Fabric diagrams of ice at Huituri off Tornio. In the upper-most of the ice the optical c-axes of the ice crystals are vertical (Fig. 2). In the lower part of the ice the c-axes are horizontal (Fig. 3). The profile of this ice is shown in Fig. 4-5.

Of all samples thin sections were made from a number of depths. These thin sections were mounted on the polaroid stage and investigated with crossed polaroids (^{5,6}).

In the first samples from Röyttä and Kuusiluoto the c-axes were vertical in all sections. But at Huituri, 6 kilometers from Röyttä, the ice showed two distinct layers; in the topmost the c-axes were oriented vertically (Fig. 2), where as they were horizontal in the lower one (Fig. 3). Farther out to sea all samples showed c-axes in the horizontal.

A close examination of the ice sample from Huituri (Fig. 4) showed it to be free from snow over the whole thickness of 31 cm. The upper part, 20.5 cm, had formed

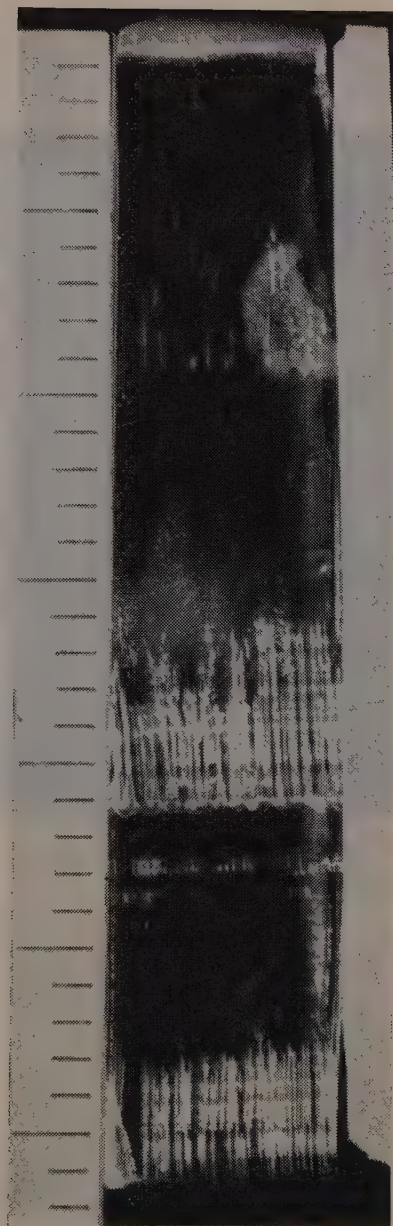


Fig. 4 — Photograph of ice from Huituri off Tornio.

during the cold spell in the end of November. After that period the weather was mild, until the temperature fell to -10°C to -15°C on the 5th December. The mild days seem to be responsible for the large air bubbles to be seen at the interface of the two layers.

The orientation of the ice crystals changes abruptly from vertical to horizontal, as a thin section extending on both sides of the interface shows (Fig. 5). The variation in salinity from day to day was not studied at this location, but without doubt the fresh river water reached the area relatively unmixed at the onset of freezing. During



Fig. 5 — Thin section of ice between crossed polaroids from Huituri area. The profile of ice is vertical. Thus the ice crystals with a vertical orientation give a white picture, the horizontal c-axes are seen dark.

the mild and windy spell the water under the ice seems to have been replaced by more saline water before the second period of freezing. The samples taken from R/V Aranda at this location (Fig. 1) in the summer show a salinity of 0.6‰, and this seems to be close to the critical value.

2. ICE IN THE LAKES

Ice samples were collected from a series of lakes. These samples were treated in the same manner as the sea samples. The hardness of the lake water was determined using standard EDTA titration.

In such lakes where the water was soft, as *e.g.* Sääksjärvi and Siikajärvi in Nuuksio, the c-axes of the ice-crystals in general were horizontal.

On the other hand, the c-axes were vertical in lakes with hard water, like Lohja lake.

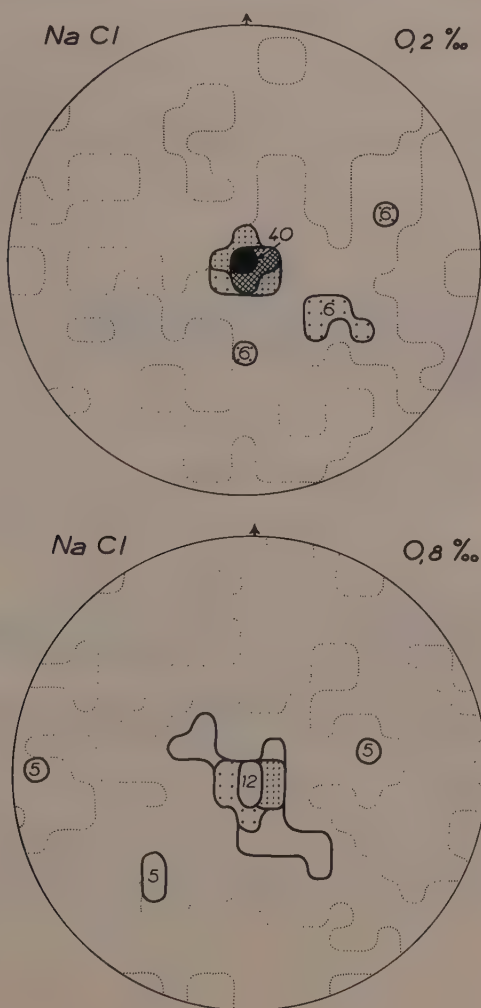
From this one might conclude that the humic colloids in the water play a greater part than expected from their electrolyte nature, in the crystal forms. In hard water,

containing appreciable amounts of calcium and magnesium ions, humic colloids tend to precipitate. In coastal regions the coagulation takes place where the river waters encounter saline waters. The bottom deposits found off Tornio during the investigation by the staff of the R/V Aranda in 1959 include some silt layers, probably formed in this way, between Kuusiluoto and Huituri.

3. LABORATORY EXPERIMENTS

To elucidate the mechanism responsible for the difference in orientation, laboratory experiments were performed.

Salt solutions containing from 0.2 to 2.0 parts per thousand of NaCl in steps of 0.2 were prepared. These solutions were frozen in a tank with well insulated sides by exposing the surface of the solution to air of -20°C . In this manner the freezing was sufficiently fast, and the presence of horizontal crystals at the sides of the tank



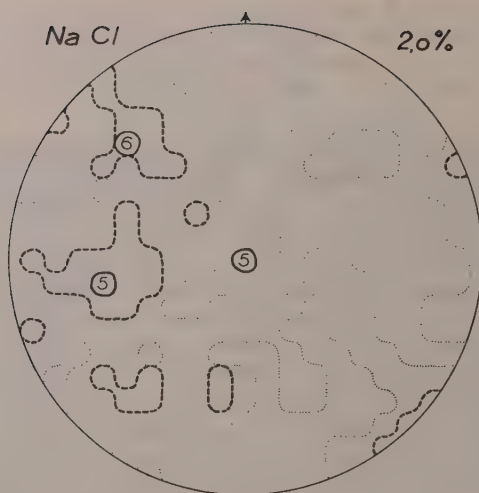


Fig. 6-8 — Fabric diagram of artificial ice. The solution contained 0.2‰, 0.8‰ and 2.0‰ of NaCl. (not 2.0% as indicated in Fig. 8).

did not appreciably influence the orientation of the ice crystals in the middle of the surface. These latter were vertical in the first stages of freezing, but the residual brine between them froze later, generally giving rise to crystals in skew positions. As the amount of NaCl increased the size and number of these latter crystals grew. The proportion of vertically oriented crystals, being 40 percent in the 0.2‰ solution (Fig. 6) dropped to 12 percent in the 0.8‰ solution (Fig. 7) and to 5 percent in the 2.0‰ solution (Fig. 8). This gives support to the supposition that the critical value is close to a salinity of 0.6‰.

As other factors than the electrolyte concentration seem to influence the phenomenon, the experiments will be extended to solutions with varying concentrations of ions and neutral particles of different size.

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EXPERIMENTAL STUDIES ON THE STRESS CONCENTRATION INDEX OF SEA-ICE

SALA, I.

SUMMARY

The theoretical stress concentration factor, α , the reduced stress concentration factor, β , and the stress concentration index, η , satisfy the equation $\eta \approx (\beta - 1)/(\alpha - 1)$. The number value of α for any hole or notch can be theoretically determined, the values of β can be experimentally determined and η , which is an approximate constant for any material, can then be calculated from the above equation. The experiments performed in Finland have given for the η of sea-ice an approximate value of about 0.

In order to experimentally measure the strength of sea ice two different methods have mainly been employed. In the first place bending tests have been performed with an in-place cantilever beam. A U-shaped channel is cut into the ice sheet and the cantilever is loaded with the force P up to the breaking point as shown in Fig. 1. The Japanese investigator TABATA (1956) and above all the others the Americans WEEKS and ANDERSON (1958) have performed experiments of this kind. In the second place ring tensile tests were performed so that a ring of sea ice is caused to fail by applying a compressive load normal to its axis as shown in Fig. 2. Of the investigators who have performed these experiments the American ASSUR (1958) at any rate must be mentioned.

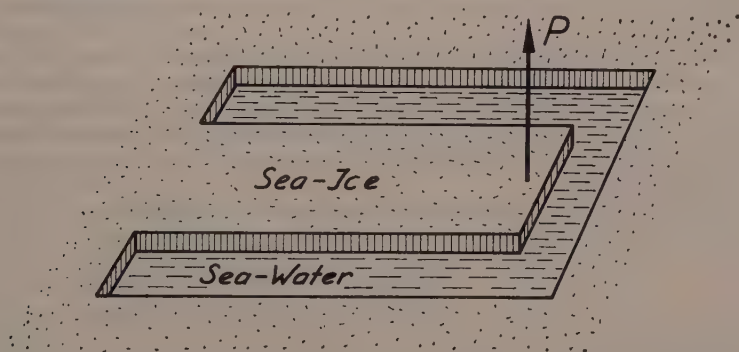


Fig. 1 — U-shaped channel in the ice sheet constitutes a cantilever beam for the bending tests (Tabata 1956).

In the beam tests the flexural strength of sea ice is computed from the formula

$$(1) \quad \sigma = \frac{6 PL}{wh^2}$$

where P is the force applied to the free end of the ice beam at failure; L is the distance from the point of application of the force to the point of breakage; w the average width of the ice beam and h the average thickness of the ice sheet. In the ring tests again the tensile strength as determined by the failure of point 2 is calculated from the formula

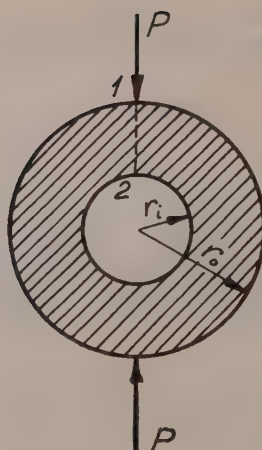


Fig. 2 — Ice ring under compression; 1-2 the critical section.

$$(2) \quad \sigma = \frac{kP}{\alpha(r_0 - r_i)L}$$

in which P is the load; r_0 the outer radius, r_i the inner radius of the ring and L the length of the ring. The constant k which is the concentration factor depends only on the ratio r_i/r_0 .

The values Obtained for σ by the ring tests have proved to be notably greater than the values obtained with the bending tests. They were in fact multiple. Even though one takes into consideration the theoretic difference which the above-mentioned two different strength values of sea ice may or can have there still remains a notable unexplained difference. In the following an attempt is made to explain this incompatibility.

Abrupt changes in the cross section of a loaded member give rise to a so-called *stress concentration*, which is given as the nominal stress multiple,

$$\sigma_{\max, \text{ theor}} = \alpha \cdot \sigma_n$$

in which σ_n is the nominal stress i.e. the average stress in the minimum section and the coefficient α , the so-called *theoretical stress concentration factor*, is considered to be dependent only on the geometrical form of the notch and on the law of elasticity chosen. It is thus possible to determine it à priori in each case separately. Different materials have, however, proved to stand in very different ways to this theoretical formula. If one experimentally determines the *reduced stress concentration factor* β from the formula

$$\sigma_{\max, \text{ exper}} = \beta \cdot \sigma_n$$

a number value is in general obtained for it which differs from the corresponding value of α and besides in the different materials in different ways. The *stress concentration index* η , which is determined by the formula

$$(3) \quad \eta = \frac{\beta - 1}{\alpha - 1}$$

has been employed and can with a certain accuracy be held as a material constant.

According to this theory of effect of notching a stress concentration arises at the built-in end of the cantilever beam used in the above bending tests. It demands the concentration factor k into equation (1) as in equation (2). On the other hand the number value of the theoretical stress concentration factor α ought not to be used for this k , as the above mentioned investigators have done, but instead the value of the reduced stress concentration factor β . But what is the magnitude of the stress concentration index η for sea ice so that the reduced stress concentration factor β would in each case be calculated.

The Institute of Marine Research in Helsinki has kindly placed some results at the disposal of the author. These results were obtained by experiments performed under the leadership of Erkki Palosuo Ph.D. outside of Helsinki on the ice of the Gulf of Finland. In spite of the fact that the tests in question were not performed specifically for this purpose some approximate values for the stress concentration index η of sea ice can, however, be obtained on their basis.

The experiments performed can be divided into two groups. One group comprises the flexural strength values obtained with two cantilever beams from otherwise the same kind of ice and under the same conditions except that the built-in end of one beam was rounded and that of the other was not. The other group comprises the ring tests which were performed under the same conditions both with a solid cylinder ($r_i = 0$) and with a cylinder with a hole in the center of it.

In the bending tests the size of the beam and the way of loading as well as the average loading rate during the tests corresponded to the experiments of the Americans. The ratio of the radius of the fillet to the width of the beam was ca. 0.15, so that according to the theory the rounded beam should have given at least $1\frac{1}{2}$ times greater values of σ than the unrounded one if both are computed from the equation (1) (see e.g. TIMOSHENKO and GOODIER 1951, p. 141). However, the experiments give the said ratio a value of about 1.03.

In the ring tests the outer radius of the ring was $r_0 = 3.8$ cm and $r_i = 0.4$ cm. According to the theory the tensile stress in the critical cross section 1-2 is then with the exception of the stress concentration appearing in point 2 very nearly constant as shown in Fig. 3 (see e.g. RIPPERGER and DAVIDS 1947). Instead of the theoretic value of about 6 the experiments give the ratio σ_0/σ , in which σ_0 was obtained for a solid cylinder from the equation (2) with the value 1 of k and σ with the cylinder with a hole in the center from the same equation and with the same value of k , the mean value of 1.143 with the standard deviation of 0.021. A consequence of this is that the number value of the stress concentration index η of sea ice probably exists in the range 0.01 to 0.04. The above explained flexural tests give the value $\eta = 0.03$.

For sea ice the stress concentration index η is, hence, very nearly zero and thus, independently of the number value of the theoretical stress concentration factor α , the reduced stress concentration factor β is very nearly 1. When this is taken into consideration a satisfactory compatibility between the results of the bending beam tests and the ring tests of the American investigators may be found.

ANDERSON and WEEKS (1958) have constructed a theoretical sea ice model shown in Fig. 4. According to this model the tensile strength of sea ice σ satisfies the equation

$$k \cdot \sigma = \sigma_p (1 - 2\sqrt{ae/\pi})$$

in which σ_p is the tensile strength of pure ice; $a = e/l_0$ (in Fig. 4); e the relative volume of brine and k the stress concentration factor in the ice bridges between the brine cylinders.

The curves shown in Fig. 5 and presented by the said investigators were obtained

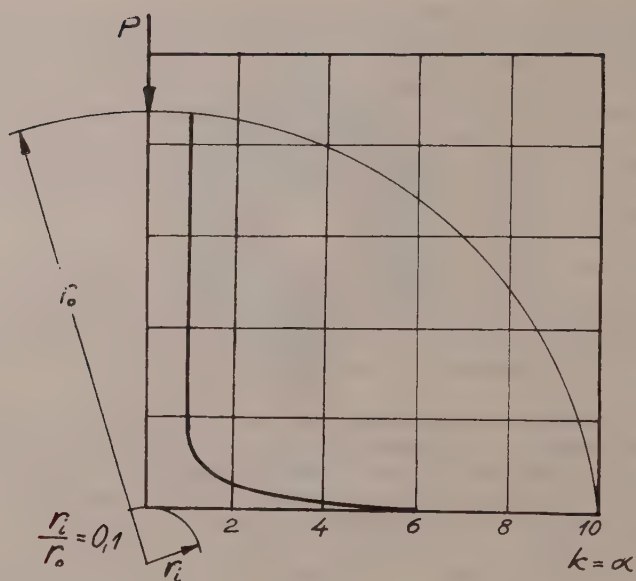


Fig. 3 — Theoretic stress distribution in the critical section 1-2 in Fig. 2 (Ripperger 1947).

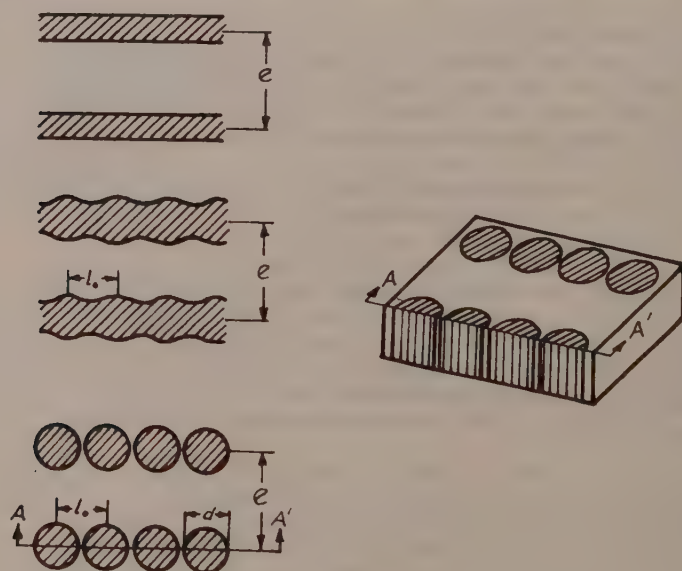


Fig. 4 — Schematic diagrams showing how a brine layer splits into cylinders (Anderson and Weeks (1958).

with the values $\sigma_p = 15 \text{ kp/cm}^2$; $a = 1$ (the dashed curve stands for $a = 2$) and $k = 3$. In the figure the short curve shows experimental data whose compatibility with the theory is apparent.

If one now takes $k = 1$, what will happen to the compatibility. In Finland in the Bay of Bothnia in the vicinity of Tornio the mean value of $\sigma = 7,5 \text{ kp/cm}^2$ has been obtained as the result of several experiments with sea ice containing very little salt so that practically no brine pockets were found in the ice. For the values of e.g. $\sigma_p = 7,5 \text{ kp/cm}^2$; $k = \beta = 1 + 0,03(3 - 1) \approx 1,1$ and $a = 2$ one practically obtains the same curve as the solid one corresponding to temperature -3°C in Fig. 5.

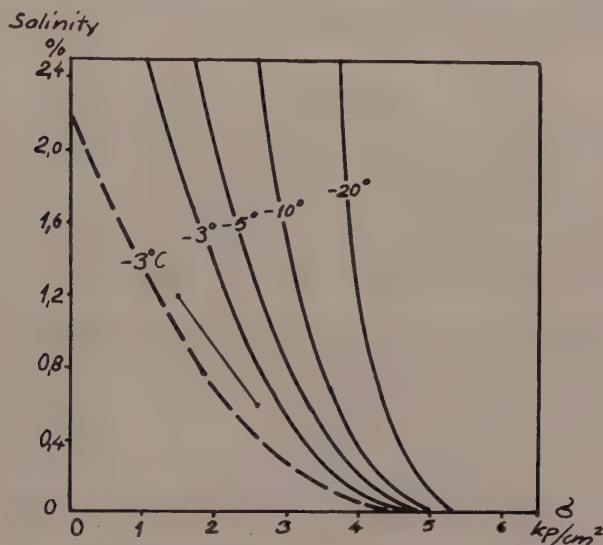


Fig. 5 — Relation between tensile strength and salinity at various temperatures (Anderson and Weeks 1958).

As it was mentioned above it was not possible for the author to perform experiment series which would have been specifically planned and vast enough for his investigations. The result presented above i.e. that the stress concentration index of sea ice is very nearly zero the author makes known with certain reservations and he hopes that the matter could be investigated more thoroughly.

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CLIMATIC CHANGE AND LONG SERIES OF ICE OBSERVATIONS AT LAKE KALLAVESI

SIMOJOKI, H.
(Hydrological Office, Helsinki)

SUMMARY

The freezing and breaking-up dates of the ice cover of Lake Kallavesi, the longest known series in Finland, are illustrated by the frequency tables. There has been a marked climatic change since the 1880's.

Ice observations have often been used for elucidation of climatic changes and cycles. Easton's research work is perhaps the best known in its kind. Ice-observation series may cover a longer period of time than observation series obtained instrumentally, such as temperature observations and others. Ice observations, provided they have been made carefully, are free from errors caused by instruments and methods.

For Lake Kallavesi (Fig. 1) there is a continuous series of freezing and breaking-up dates of ice cover beginning in late autumn 1833. The observations bear upon the open lake outside Kuopio. The greatest depth of the lake is 48 meters (Fig. 1). For a long

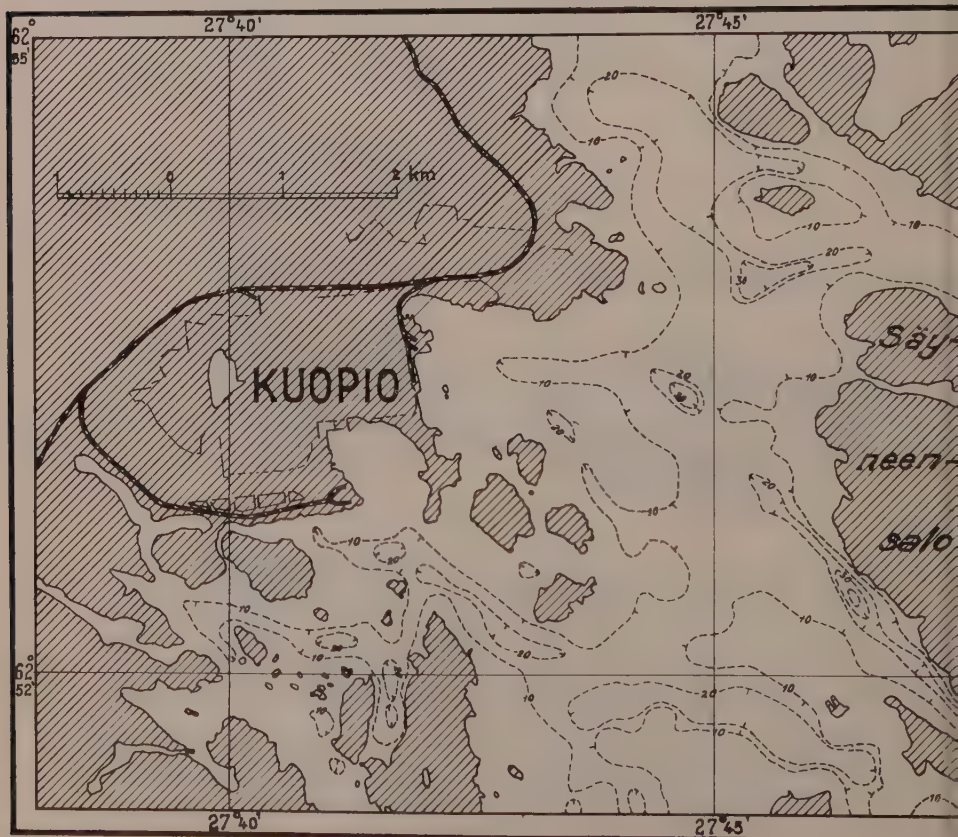


Fig. 1

TABLE 1

The freezing dates of the Lake Kallavesi. Winters 1834-1957.

November	3	1865						
	4							
	5							
	6	1853	1881					
	7	1882	1922					
	8	1895						
	9	1957						
	10	1942						
	11	1857	1928					
	12							
	13	1883	1920					
	14	1835	1842					
	15	1909						
	16	1841	1845	1849	1862	1868		
	17	1859	1876	1903				
	18	1869	1877					
	19	1836	1934					
	20	1861	1872	1892				
	21	1843	1851	1886	1888	1916		
	22	1837	1839	1867	1894			
	23	1889	1891					
	24	1855	1856	1871	1902	1905	1926	
	25	1844	1850	1874	1880	1885	1900	
	26	1860	1956					
	27	1840	1858	1875	1897	1910	1923	1953 1958
	28	1863	1901					
	29	1924	1943					
	30	1898	1911					
December	1	1847	1866	1879	1918	1927	1938	
	2	1893	1904	1946	1955			
	3	1907	1908	1925	1952			
	4	1852	1873					
	5	1899						
	6	1932	1941					
	7	1884	1919					
	8	1896						
	9	1870	1914	1947				
	10	1838	1912	1949				
	11							
	12	1929	1940	1945				
	13	1834	1846	1948				
	14	1906	1931					
	15	1890						
	16	1921						
	17	1913	1939					
	18	1864	1915	1917				
	19	1854	1887					
	20	1954						
	21	1848						
	22	1936						
	23							
	24							
	25	1878	1935	1950	1951			
	26	1937						
	27							
	28	1944						
	29							
	30							
	31							
January	1							
	2							
	3							
	4							
	5							
	6							
	7	1933						
	8							
	9							
	10							
	11							
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	24							
	25							
	26							
	27	1930						

The breaking-up dates of the Lake Kallavesi. Winters 1834-1957.

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The observations of winters 1934-1957 are shown in Tables 1 and 2. The records are grouped according to calendar years. The years refer to the spring-part of the winter. Thus, if freezing has occurred before the 1st of January the year refers to the spring-part of the actual winter.

From the values in Table 1 it will be seen that the arithmetical mean of freezing date is November 30. Its dispersion is $\zeta_Z = 14$ days. The mode falls on November, 27, and the frequency curve is positively skewed. During the 124 years under observation the earliest freezing had occurred in 1864, November, 3, and the latest in 1930, January, 27. The difference in time between these is 85 days. Greatest departures are due to the delay in freezing, owing to the fact that the heat supply of the water must decrease to a certain value before the freezing can take place. Provided this thermal situation has been arrived, and the weather remains relatively warm at the same time, the freezing is delayed.

For the arithmetical mean of breaking-up date of ice cover (Table 2) the date May, 18, is obtained. The dispersion is $\zeta_A = 9$ days. Thus it is a considerably more regular phenomenon than freezing. The earliest date of breaking-up has been in 1921, April, 20, and the latest in 1867, June, 17. The range of variation has been 58 days. The mode falls on May 24. From the general structure of the Table it is seen that the frequency curve is negatively skewed.

For establishing the alterations in climate the records are divided in two successive groups of 62 winter each. For these the following means are obtained:

	1834-1895	1896-1957
Freezing	Nov. 23	Dec. 3
Breaking-up	May 22	May 14

A survey of these dates indicates that the freezing in the latter period is delayed on an average of 10 days and the breaking-up of ice cover occurs 8 days earlier during the same period.

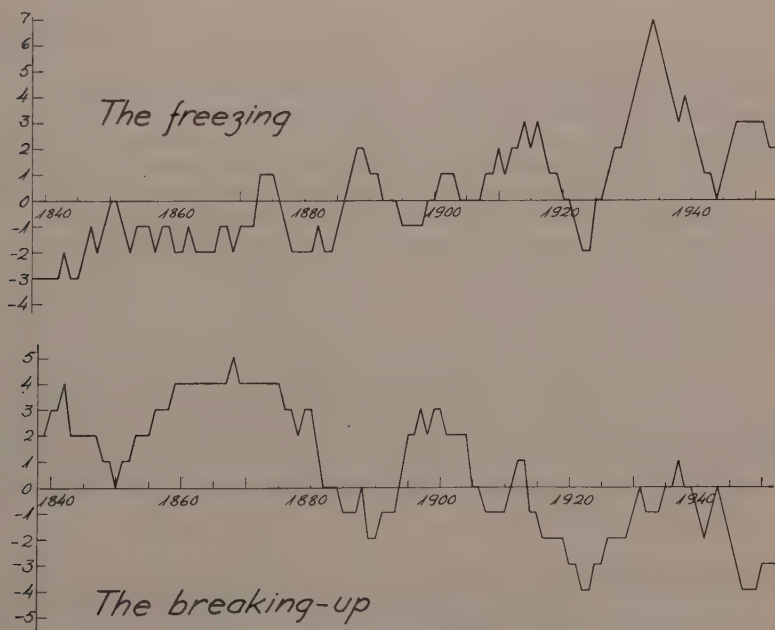


Fig. 2

Following proceedings will be introduced for obtaining a more detailed comprehension on this matter:

The arithmetic mean of the whole observation series of the freezing dates is denoted Z_m and the dispersion ζ_Z . The number of cases over a period of 10 years, when the freezing had occurred later than $Z_m + \zeta_Z$ is marked m and the number of cases over the same period, when the freezing had occurred earlier than $Z_m - \zeta_Z$ is denoted n . Furthermore, the differences $m - n$ are computed for as many 10-years periods as possible, e.g. for the periods 1834-1843, 1835-1844 etc.

The same method will be applied to the dates of breaking-up. The number of cases over the same period when the breaking-up had occurred later than $A_m + \zeta_A$ is denoted m and the number of cases when the breaking-up had occurred before $A_m - \zeta_A$ is denoted n . The course of the differences $m - n$ gives an idea of the possible variations.

The differences $m - n$ for the freezing and breaking-up are illustrated in Fig. 2. It will be seen that the freezing and breaking-up dates have a clear secular course. According to the Figure, there are also fluctuations of a duration of some years, which represent the nature of the general alteration. These results are in good agreement with other climatological elements, e.g., the temperature in Helsinki has a similar course.

GENERAL PROPERTIES OF ARCTIC SEA ICE ¹

E.R. POUNDER and P. STALINSKI

RÉSUMÉ

Des témoins de glace ont été extraits de quelques endroits du Détroit de Barrow. De ces témoins on a préparé les profils de la salinité, la densité, la structure cristallographique, la température de la glace et le module de Young. Les comptes-rendus de la profondeur de la glace et des conditions météorologiques étaient à la portée et on a pu calculer la date de congélation de quelques couches de la glace. La conductibilité thermique moyenne, telle que calculée, est de 2.14×10^{-3} c.g.s. La salinité varie entre 9 et 4 pour-mille. On a trouvé une bonne corrélation entre les augmentations abruptes de la salinité et celles du taux de gel. Le poids spécifique relatif moyen est de 0.94. Les photographies des couches de la glace montrent que la structure principale de cette glace est formée par des cristaux verticaux, très longs, sans vides d'importance.

SUMMARY

Ice cores were extracted at several locations in Barrow Strait. From the cores, observations were made leading to profiles of salinity, specific gravity, crystal structure, ice temperature and Young's modulus. From available meteorological and ice thickness records the growth rate of the ice was known, permitting calculation of the date of formation of any layer of ice. The average thermal conductivity of the ice was calculated as 2.14×10^{-3} c.g.s. units. Salinity values ranged from 9 to 4 ‰. Sharp increases of salinity with depth correlated well with increases in the freezing rate. The average specific gravity was 0.94. Crystal photographs showed that most of the ice cover consisted of very long vertical crystals with no appreciable voids.

1. INTRODUCTION

In the spring of 1959 an investigation was made in the vicinity of Cornwallis Island of some of the properties of annual, arctic sea ice. The present paper relates to general weather conditions and the observations on the resulting factors of salinity, specific gravity, growth rate, and crystal structure. A later paper (Pounder and Stalinski 1960) discusses results obtained on the elastic moduli.

Fig. 1 illustrates the region in which observations were made. A base camp was established at $74^{\circ} 43.5' \text{ N}$, $96^{\circ} 27' \text{ W}$ and occupied from about April 22 to May 10. At this location the water is about 700 feet deep and the ice cover at the time was about 6.5 feet thick. A number of cores were extracted at base camp, using a power-operated SIPRE ice corer, and tested. These will be referred to as Cores BC1, BC2, etc. A party travelled south-west on the track shown from April 30 to May 3. Cores were extracted at the sites marked on Fig. 1. At each site two cores were drilled at points within a foot of each other. These cores are referred to as 6A, 6B, 9A, 9B, etc., the number giving the site location. The A cores were sectioned and used to obtain specific gravity, elastic constant, and crystal structure profiles. The B cores were sectioned and melted to find salinity profiles.

(¹) Une contribution du projet de la recherche sur la glace, Département de Physique, Université McGill, Montréal, Canada.

(¹) Contribution from the Ice Research Project, Department of Physics, McGill University, Montreal, Canada.

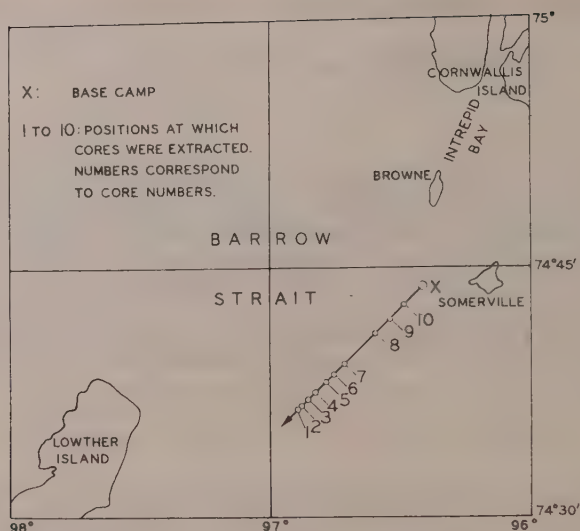


Fig. 1 — Location of ice observations.

2. SALINITY AND SPECIFIC GRAVITY

Each core used to obtain a salinity profile was cut into sections 2.5 to 3.5 inches long, sealed in polythene boxes, and allowed to melt. Twelve salinity profiles were obtained in this way, the salinity of each melt being measured with a hydrometer and corrected to 15°C. Four typical profiles are shown in Figs. 2 and 3. The date after each core number is the date it was extracted from the ice. The considerable variation in the profiles almost certainly represents a real variation in the ice from one location to another, as the probable error of the hydrometer readings should not be more than 0.5‰.

In an attempt to obtain a smoothed salinity curve representative of the entire region, the data were classified in two-inch layers, e.g. the salinities of all samples which included one or more inches of the layer between 12 and 14 inches in depth were listed. For each layer the average of all the values obtained in this way was calculated. Fig. 4 shows a smoothed salinity curve based on these average values. By integrating this curve graphically the average salinity of the ice was found to be 5.1‰.

Specific gravity profiles were obtained by taking core sections of cylindrical form, measuring their dimensions, and weighing them. Errors in weighing were negligible but uncertainties in measuring the volume of the samples gave rise to possible errors of 2 to 3%. All results obtained for specific gravity were within 2% of their average and the random nature of the profiles when plotted showed that with the method used significant variations in the specific gravity could not be detected. The average value was 0.943 ± 0.02 .

3. ICE THICKNESS, GROWTH RATE AND TEMPERATURE REGIME

The thickness of the ice was measured in 24 drill holes at 11 different locations. The average thickness was 84.7 inches, the maximum thickness was 97 inches, and the minimum 74.5 inches. There was no trend in thickness along the 10-mile strip surveyed.

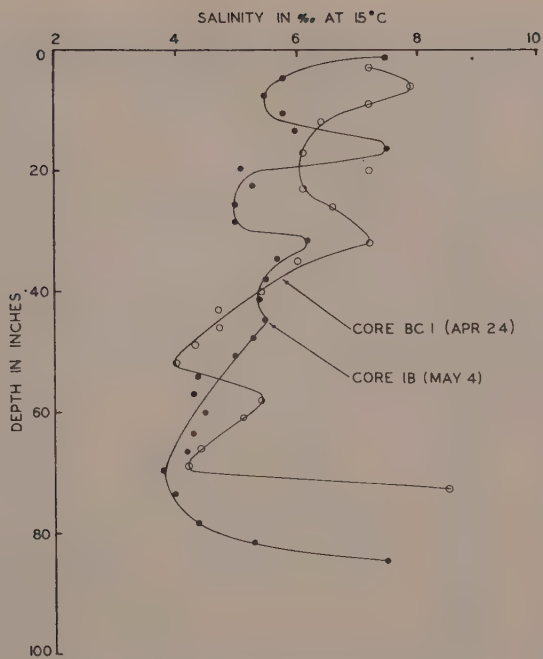


Fig. 2 — Salinity profiles of cores BC1, 1B.

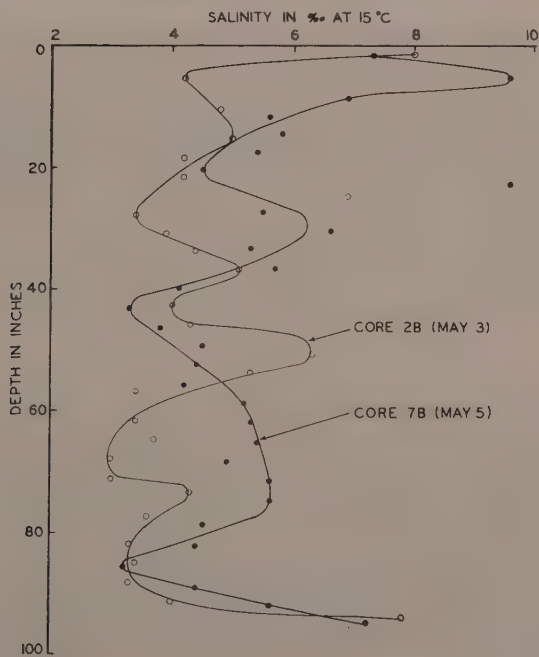


Fig. 3 — Salinity profiles of cores 2B, 7B.

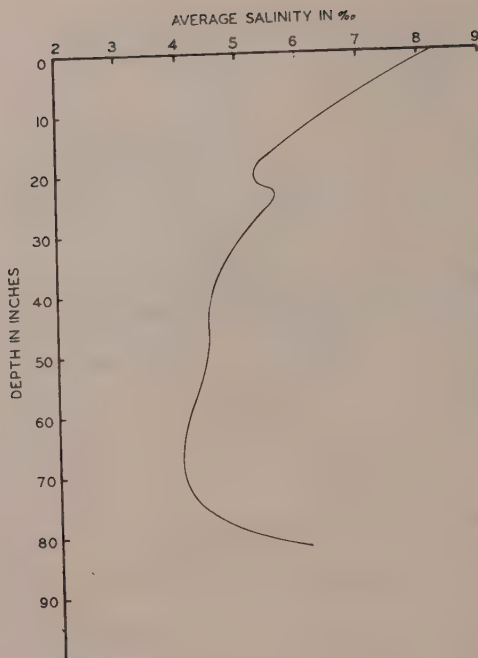


Fig. 4 — Average salinity profile.

Meteorological data at the RCAF base at Resolute and information on the thickness of the ice in Resolute Bay (about 26 miles east of base camp) were supplied by the Joint Weather Station, Resolute Bay. The first continuous ice cover started to form on Sept. 29, 1958. The thickness of the ice in the bay was measured weekly, and is shown in Fig. 5, plotted against the square root of the freezing exposure E , in fahrenheit degree-days below 28.7°F . In calculating E , the mean of the daily maximum and minimum temperatures was taken as the mean temperature. Exposure was cumulated daily.

Pounder and Little (1959) show that the simplest expression for the ice thickness h is

$$h = \left(\frac{2kE}{Ld} \right)^{1/2} \quad (1)$$

where k , L , and d , are respectively the thermal conductivity, latent heat of fusion, and density of the ice. Fig. 5 shows that this linear relation between h and \sqrt{E} applied quite well for the growth of this ice sheet up to about March 1. After this the ice thickened more rapidly than the linear rate, although one would have expected it, if anything, to grow more slowly because of the increased heat contribution by solar radiation, which was disregarded in deriving equation (1).

Using the linear part of the graph, an average thermal conductivity can be calculated if L and d are known. There is some uncertainty about the proper value of latent heat to take. Sea ice growing from brine releases less heat than pure ice does because of the brine trapped in the ice. Also heat must be continually extracted from

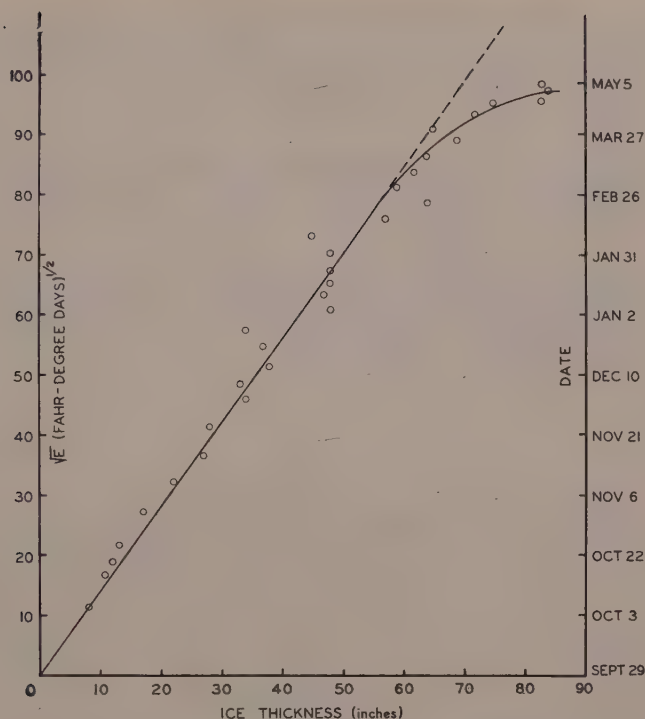


Fig. 5 — Ice growth as a function of freezing exposure and date.

the upper layers of an ice sheet as the freezing interface moves down, the quantity depending both on a specific heat term and on a term which represents the increasing concentration of the brine cells. The Malmgren (1927) formula,

$$L = L_p \left(1 - \frac{S_i}{S_w} \right) \quad (2)$$

where L_p is the latent heat of pure water and S_i and S_w are the salinities of the sea ice and sea water, gives a value of the latent heat which allows for at least part of these effects. Taking S_i as 5.1‰ and S_w as 31.7‰ (value measured at 4 m depth with a Nansen bottle),

$$L = 79.7 \left(1 - \frac{5.1}{31.7} \right) = 66.9 \text{ calories/gram.}$$

The slope of the growth curve gives

$$\frac{E}{h^2} = \left(\frac{100}{71} \right)^2 \text{ fahr. degree-days inch}^{-2} = 1.48 \times 10^4 \text{ c.g.s. units.}$$

Substituting in (1),

$$k = \frac{Ld}{2} \frac{h^2}{E} = \frac{66.9 \times 0.943}{2 \times 1.48 \times 10^4} = 2.14 \times 10^{-3} \text{ c.g.s. units.}$$

This value of the average thermal conductivity of the ice sheet is lower than those frequently quoted by others, such as 5×10^{-3} c.g.s. units. The insulating effect of the snow cover is neglected in the calculation, but the snow should not have had a major influence (except to reduce radiation effects greatly) since it never reached a foot in thickness, and most of the winter consisted of only six inches of hard, wind driven cover.

Another indication of the abnormally low growth rate up to March 1 is obtained by applying the empirical formula of Assur (1956). For the growth of ice he gives

$$h \text{ (in inches)} = 1.06 \alpha \sqrt{S} \quad (3)$$

where S is the freezing exposure in Fahrenheit-degree days below 32°F and α is an empirical coefficient dependent on snow cover and type of ice. The freezing exposure S is somewhat greater than E . On March 1 the value of E was about 6600 and the corresponding linear value of h was 57.5 inches.

$S = E + 3.3 \times \text{no. of days between Sept. 29 and Mar. 1}$
 $\approx 7100.$

$$\text{From (3), } \alpha = \frac{57.5}{1.06 \sqrt{7100}} = 0.64.$$

Assur quotes values of α between 0.70 and 0.75 for Arctic sea ice.

Fig. 6 is a plot of the daily mean temperature at Resolute Bay from Sept. 1 to

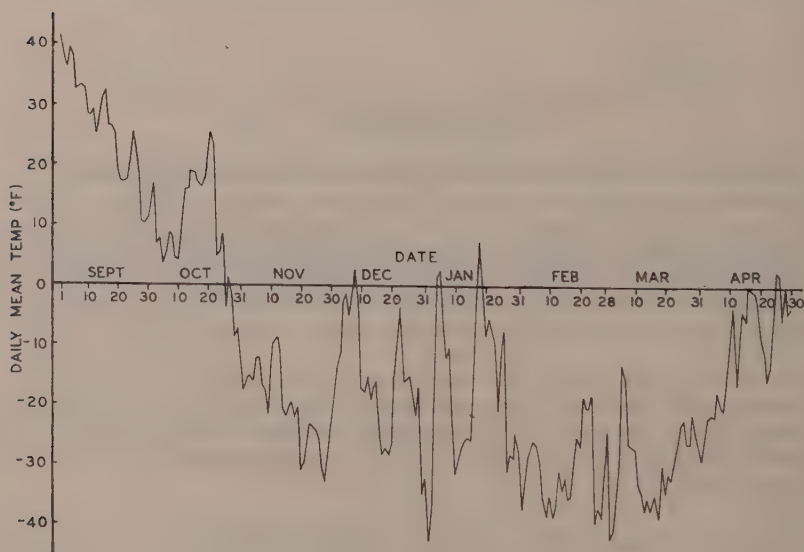


Fig. 6 — Mean daily temperatures at Resolute Bay—winter of 1958-59.

Apr. 30 and Fig. 7 shows the daily maximum and minimum temperatures, also at Resolute, during the period of the expedition.

The weather and ice thickness data obtained at Resolute Bay permit approximate dating of the layers of the ice samples obtained on this expedition. Fig. 5 has an additional scale at the right giving the date corresponding to each value of E . This

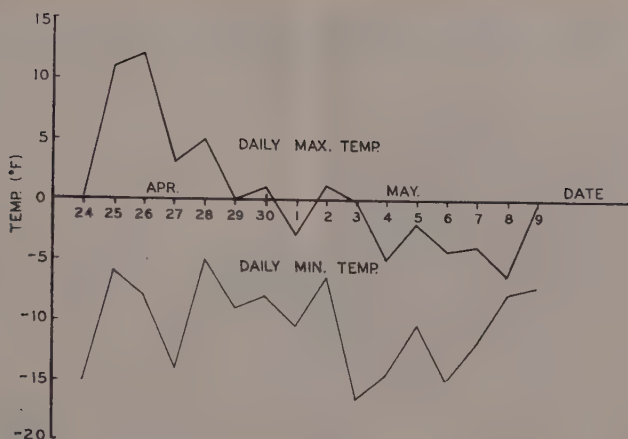


Fig. 7 — Daily temperatures at Resolute Bay—April-May 1959.

date scale is naturally nonlinear. As an example of its use, the mean salinity curve of Fig. 4 has one main anomaly, the peak at a depth of 23 inches. From Fig. 5, this layer was frozen about Nov. 9. Fig. 6 shows that this increase in salinity resulted from the rapid freezing following the markedly lower temperatures starting Oct. 29. Analysis of a highly simplified model of the ice situation at this time is given by Pounder (1960). This analysis shows that a lag of the order of a few days between the drop in temperature and the rapid increase in thickness and salinity is reasonable. Similarly, Fig. 4 suggests a slight increase of salinity at depths of 45-50 inches. Peaks at this depth show clearly on several of the individual salinity profiles. This probably resulted from the extremely cold temperatures between Dec. 28 and Jan. 3 although the formation of the 50-inch layer can be dated approximately as taking place on Jan. 31.

4. CRYSTAL STRUCTURE

After other tests on the cores were completed a number of cores (BC2, 1A, 2A, 7A, and 9A) were analyzed for crystal structure.

Three-inch sections were chosen at intervals in each core, and from each section horizontal and vertical slabs were cut. These slabs were melted down on a warmed metal plate to thickness of from 1 to 2 mm and photographed in a simple polariscope. In each photograph the vertical section (on the left) is oriented with the top of the ice upward. The scale can be obtained from the semi-circular horizontal sections which are three inches in diameter. Figs. 8 to 10 show selected photographs of parts of five of the cores.

All of the crystal analyses gave similar results and showed few surprising features. The general pattern is one of small, randomly-oriented crystals in the surface, with a transition layer (Perey and Pounder 1958) at a depth of about half an inch. Below this the structure rapidly changes to one of long vertical crystals with presumably horizontal c-axes. Crystal size tends to increase with depth until 3-inch cores from the lowest parts of the ice cover frequently contain only two or even one crystal. The section from 90.7 to 93.7 inches in Fig. 8 is an example of a single crystal with only very minor inclusions. A thin skeleton layer was observed on the bottom of



Fig. 8 — Crystal structure of core 7A. The depths of the sections in inches from the top surface of the ice cover, are: (top left) 0-3'', 5.2''-8.2'', 27.2''-30.2'', (top right) 90.7''-93.7'', 93.7''-96.7''.

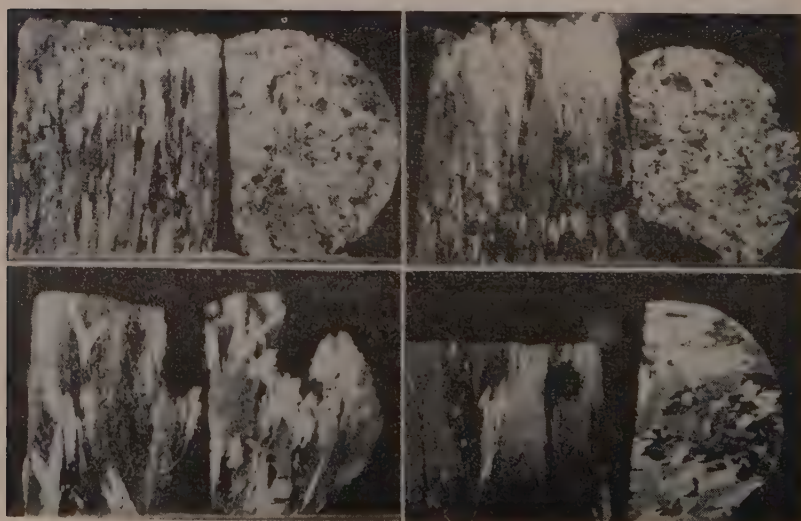


Fig. 9 — Crystal structure of cores BC2 (left), 9A (right). Depths are: on the left 1''-4'' and 71''-74'' and on the right 0''-3'' and 80.5''-84.5''.

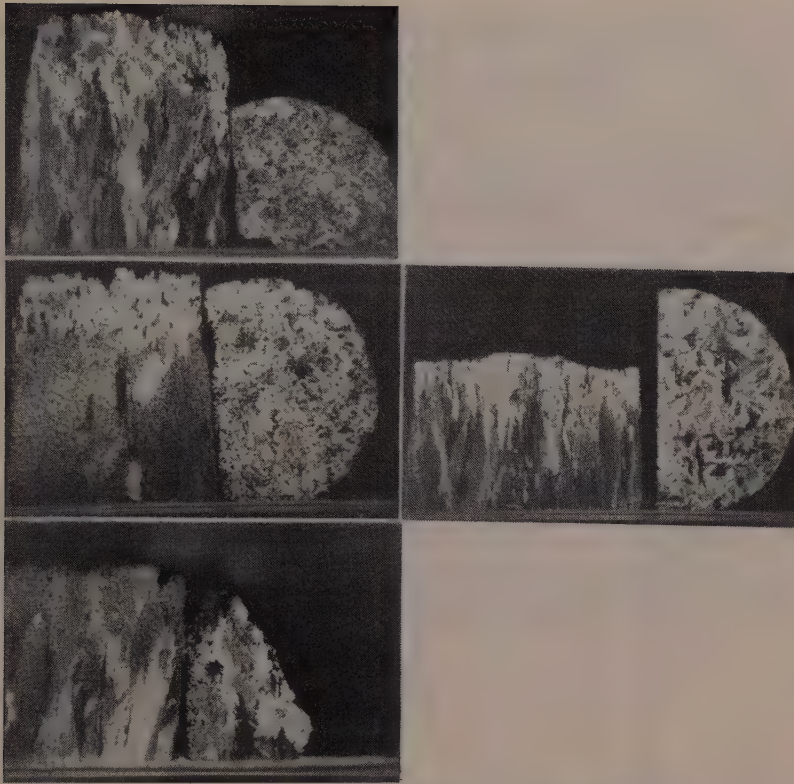


Fig. 10 — Crystal structure of cores 2A (left), 1A(right). Depths are: on the left 5.5'' to 8.5'', 27''-30'', 84.5''-87.5'' and on the right 17''-20''.

most cores when they were extracted. This layer was never more than half an inch thick and does not show in the photographs as it was melted in preparing the thin sections.

The lowest section of each core (see the bottom sections of cores 7A, BC2, 9A, and 2A in the Figs.) has a less ordered structure than those immediately above it. This suggests that the ice on first freezing has a number of randomly oriented small crystals and that metamorphosis to a more ordered arrangement takes place fairly rapidly in this ice which remains for a long period of time at a temperature just below its freezing point. Some other, finer details are observed on comparing Fig. 8 with the salinity profile of the immediately adjacent core 7B (Fig. 3). The maxima in salinity at 5 inches and 30 inches have already been interpreted as resulting from more rapid freezing, which would favour the starting of new small crystals. The 5.2 to 8.2-inch and 27.2 to 30.2 inch photographs bear this out. Similarly, Fig. 10 shows rapid freezing at 27 to 28 inches in core 2A and at 17 to 18 inches in core 1A. Fig. 3 shows an anomalously high salinity at 25 inches in core 2B and a sharp peak in salinity at 16 inches in core 1B.

5. TEMPERATURE PROFILES

On three occasions, temperature profiles in the ice cover were measured by drilling and extracting a core. As rapidly as possible, a radial hole was drilled into the

core and the bulb of a toluol-filled glass thermometer was inserted to the centre of the core. After a steady reading was obtained (in about half a minute), the process was repeated a few inches along the core. Measurements were started at the bottom of the core and a complete profile obtained in about 20 minutes. Fig. 11 shows the results. Individually, the curves appear not unreasonable. The curve for April 29 can be interpreted as a linear gradient with a surface temperature of -25°C , but with the top 20 inches modified by the increasing air temperature. Similarly, the curve of May 4 can be seen to show the same warming trend and also the subsequent drop in air

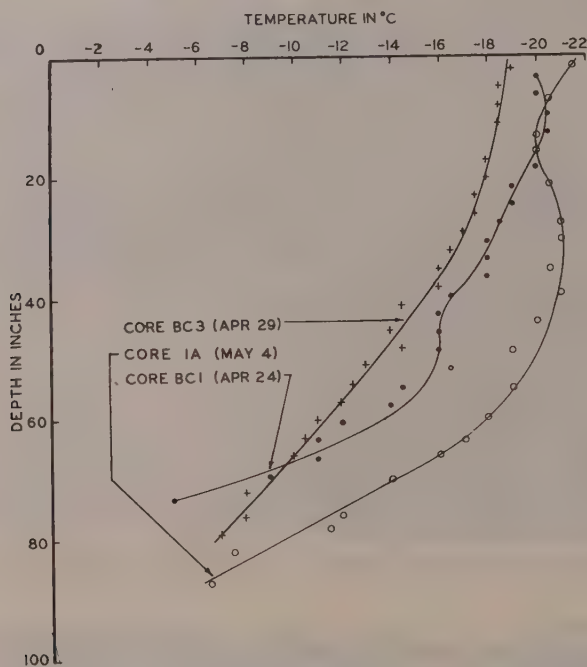


Fig. 11 — Temperature profiles in ice cover—April 24, April 29, and May 4, 1959.

temperature between April 29 and May 4. However, if the linear portions of the three curves are extrapolated to the surface they give such discordant temperatures as -45°C , -25°C , and -50°C . Apparently there was some systematic error in the lower measurements and the method can be considered as only qualitative.

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ELASTIC PROPERTIES OF ARCTIC SEA ICE ¹

E.R. POUNDER and P. STALINSKY

RÉSUMÉ

Les vitesses de la propagation du son dans la glace sont mesurées par les observations des temps de passage des pulsations longitudinales et transversales. Les transducteurs de titanate de baryum sont montés aux deux bouts de cylindres coupés dans des témoins de glace de mer et ayant un diamètre de 7.5 cm et une longueur de 15 cm, environ. De ces deux vitesses on peut calculer le module de Young (E) et le rapport de Poisson (σ). On montre que la valeur moyenne de E de cette couverture de glace est de 8.42×10^{10} dynes-cm⁻² et qu'elle diminue dans les couches supérieures et inférieures où la salinité est plus forte. La relation entre E et la salinité S (en pour-mille) est donnée par l'équation :

$$E = (9.75 - 0.242 S) \times 10^{10}, \text{ c.g.s.}$$

SUMMARY

The speed of sound in ice was measured by observations on the transit times of pulses of longitudinal and transverse waves. Barium titanate transducers were frozen to both ends of cylinders, about 15 cm long by 7.5 cm in diameter, cut from sea-ice cores. From the two velocities, Young's modulus (E) and Poisson's ratio (σ) can be found. It was shown that the average value of E for this ice cover was 8.42×10^{10} dynes cm⁻², and that its value was lower for the top and bottom layers where the salinity was higher. The relation between E and the salinity S (in ‰) was given by the equation

$$E = (9.75 - 0.242 S) \times 10^{10} \text{ c.g.s. units.}$$

1. INTRODUCTION

A previous paper (Pounder and Stalinski 1960) has described an expedition in the spring of 1959 to Barrow Strait, west of Cornwallis Island, and gave an account of the weather during the winter of 1958-59 and of the results obtained for such dependent variables as salinity and crystal structure of the ice cover. A major objective of ice work on this expedition was to attempt to measure the elastic properties of the ice and to look for correlations between them and the other variable properties of the sea ice. The elastic moduli of sea ice have been measured comparatively rarely, and with widely varying results.

The method used to obtain the elastic properties consisted of measuring transit times for pulses of supersonic frequency through sections of a core. If both longitudinal and transverse wave velocities are found, then Young's modulus E and Poisson's ratio σ can be calculated. The apparatus used was designed and constructed by Dr. M.P. Langleben. Initially, an extended series of laboratory experiments with the equipment had been planned. However, time permitted only a few laboratory tests in advance and one brief field trial on salt water ice samples, so that the method was unfortunately relatively untried.

(¹) Une contribution du projet de la recherche sur la glace, Département de Physique, Université McGill, Montréal, Canada.

(¹) Contribution from the Ice Research Project, Department of Physics, McGill University, Montreal, Canada.

2. EXPERIMENTAL METHODS AND RESULTS

The principle used is illustrated in the block diagram of Fig. 1. A DuMont Type 326 time-delay generator initiates a series of pulses which are amplified by a hydrogen thyratron pulse amplifier and used to excite a barium titanate crystal transducer, cemented (by freezing it with a few drops of fresh water) to one end of the ice sample. A similar crystal at the other end of the sample receives the pulses and after suitable amplification they are applied to the vertical plates of a cathode ray oscillograph. Each pulse from the generator also travels through a variable time-delay line to trigger the horizontal sweep of the C. R. O. By changing the time delay, the pattern on the scope can be shifted with respect to the graticule in order to measure the transit time of the pulse, and hence the velocity of the supersonic waves in the ice.

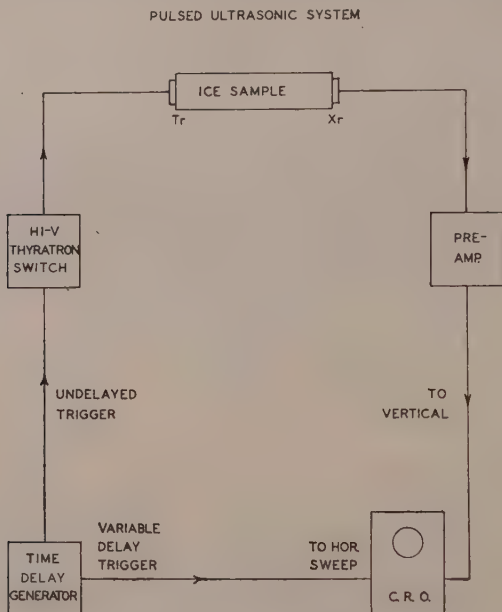


Fig. 1 — Pulsed system for measuring wave velocities in ice.

The barium titanate crystals used had a natural frequency of oscillation of 500 Kc per second and the crystals found most effective were those designed to produce primarily transverse vibrations so that they excited strong shear waves. It was found that these crystals also gave quite detectable longitudinal waves. The pulse repetition frequency was 60 cycles per second and the pulse length 5 microseconds.

The ice samples used were sections of the 3-inch (7.5-cm) core about 6 inches long. The compressional wave velocity measured was about 3750 m sec^{-1} , so that the wavelength of the longitudinal waves was 0.75 cm or one tenth of the core diameter. The shear wave velocity is appreciably lower so that the wavelength of the shear waves was even smaller. Under these circumstances the ice sample can be considered to be of infinite extent and the relevant equations for the two velocities are

$$c_B = \sqrt{\frac{E}{d} \frac{(1 - \sigma)}{(1 + \sigma)(1 - 2\sigma)}} \quad (1)$$

and

$$c_S = \sqrt{\frac{E}{d} \cdot \frac{1}{2(1+\sigma)}} \quad (2)$$

where c_B , c_S are the longitudinal (or bulk) and shear velocities, and d is the density of the ice. The ratio of these equations gives

$$\frac{c_B}{c_S} = \sqrt{\frac{2(1-\sigma)}{1-2\sigma}} \quad (3)$$

permitting calculation of Poisson's ratio from measurements of the two velocities.

Fig. 2 is a sketch of the typical pattern seen on the oscilloscope. The start of the input pulse is marked at I and the start of the received compressional-wave pulse at C . The transit time for compressional waves was easily measured but the start of the shear wave is more difficult to pick out. Apparently the shear waves were reflected strongly. In the field observations, the peak of the first major negative shear displacement (marked S in the figure) was selected.

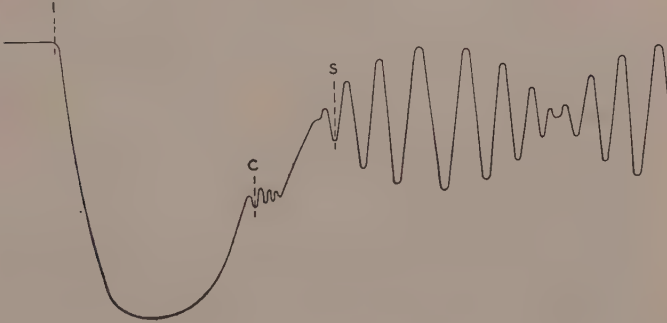


Fig. 2 — Typical wave form of transmitted pulse of shear and compressional waves

The temperature of the ice being measured is not exactly known. Cores were stored for two days or more outdoors at ambient temperatures of -18°C to -22°C . For each test the section of the core was brought into a heated room at about 20°C . The density and sonic pulse measurements were completed in about eight minutes. In view of the consideration on heat conduction in cylinders in Pounder (1960), the temperature of the bulk of the ice cannot have increased by more than one or two degrees. Since timing the start of the pulse train gives the highest velocity, which is characteristic of the coldest part of the ice, it is reasonable to conclude that the values obtained refer to a temperature not more than two degrees above the outside air temperature. Northwood (1947) has shown that the velocity of longitudinal waves changes very slowly with temperature, so that it appears that no appreciable error would be made in considering that the wave velocities were measured in all cases at a temperature of -20°C . Table I shows in detail the data and calculations for core 7A, which was quite a typical ice sample. The column marked depth gives the location of the ice section below the surface. The columns σ and E give the values calculated from equations (1) and (3). The column E' is discussed below.

For fresh-water ice, the best values of σ and E are probably those of Northwood (1947), respectively 0.33 and 9.8×10^{10} dynes cm^{-2} . For sea ice, various reports show a wide range of values. Anderson (1958) reviews dynamic tests and gives E values

from 1.7 to 7.28×10^{10} dynes cm^{-2} . Static methods (see Tabata (1958) for example) give an even wider spread. For the heavy, cold, Arctic sea ice being studied here one would expect values of σ and E approaching those of fresh-water ice. One would also expect reasonable consistency in the observed values because of the uniformity of temperature and the accurate reproducibility of the sonic method employed. The calculated values of σ are almost certainly too high and the spread (from 0.27 to 0.43) is too great. The average value was 0.386 . Averaging all the E values gives 6.37×10^{10} dynes cm^{-2} .

The calculated profiles of σ and E showed large and apparently quite erratic variations, and it seems apparent that the method of timing the shear pulses was inadequate. Hence the results for Young's modulus have been recalculated assuming constant values of $\sigma = 0.33$ and density $d = 0.943 \text{ gm cm}^{-3}$. The recalculated values for core 7A are shown as E' in the last column of Table I. The values are much more consistent and show a dependence of E' on salinity, as would be expected. Two very low values, caused presumably by errors in experimental observations, have been discarded. The average of the remaining 90 values is $E = 8.42 \times 10^{10}$ dynes cm^{-2} . The top and bottom sections of each core were then considered separa-

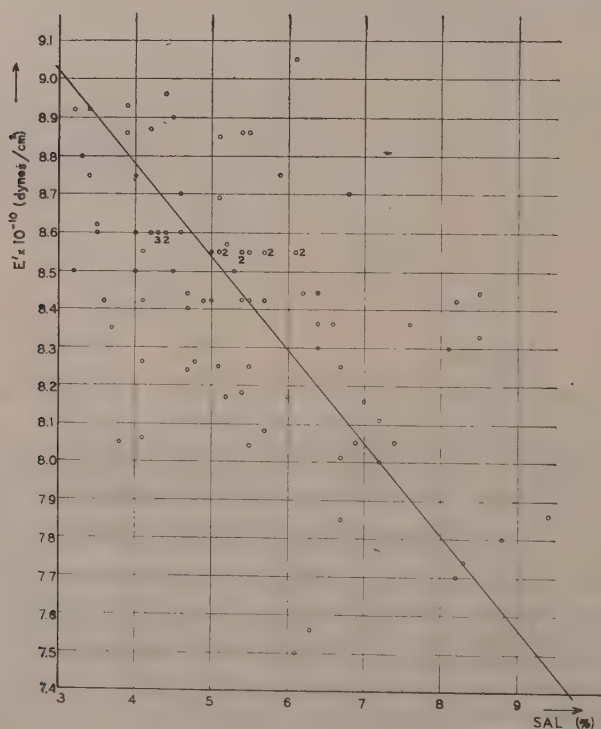


Fig. 3 — Variation of Young's modulus with salinity.

tely, since these are the regions of highest salinity. The average for the twelve surface sections is 8.03 and for the bottom layer 8.31×10^{10} c.g.s. units. Disregarding these, the remaining 66 core sections give an average value of Young's modulus of 8.51×10^{10} dynes cm^{-2} . This is considered the best figure obtained for Young's modulus for the bulk of the ice cover. The average deviation of the 66 values from this average is 2.4% and the extreme deviation is 12% .

To investigate further the correlation between Young's modulus and salinity, the average salinity S for each of the 90 core sections was found from the relevant salinity profiles, and E_c was plotted against S (Fig. 3). The points were analysed by the least-squares method to give the best linear fit shown. The equation of the line is $E' = (9.75 - 0.242 S) \times 10^{10}$ dynes cm^{-2} , with S in ‰. The extrapolation to a value of 9.75×10^{10} dynes cm^{-2} for Young's modulus for pure ice is gratifying. The correlation coefficient between E' and S is -0.78 , indicating a fairly high validity for this equation.

Mr. T.A. Harwood of the Defence Research Board and Lieut-Commander J.P. Croal of the Royal Canadian Navy supplied valuable experience in Arctic conditions. The assistance of the Royal Canadian Air Force in supplying air transportation is gratefully acknowledged. This work was supported by the Defence Research Board of Canada through D.D.P. Contract G.C. 69-900109.

TABLE I
Elastic Properties

Depth (inches)	Length (cm)	Delay in μsec		$\frac{c_B}{c_S}$		$\frac{c_B}{c_S}$	σ	d (gm/ cm^{-3})	$\frac{E}{E'}$	
		Compress.	Shear	(cm/sec $\times 10^{-10}$)		$\frac{c_B}{c_S}$			(dynes cm^{-2} $\times 10^{-10}$)	
0-4	9.8	28.0	72.1	35.0	13.6	2.57	0.411	0.933	4.93	7.74
5-11	15.7	43.6	92.2	36.0	17.0	2.12	0.357	0.938	7.35	8.16
11-17	15.6	42.7	98.8	36.6	15.8	2.32	0.386	0.940	6.46	8.42
17-23	15.0	41.5	98.0	36.1	15.3	2.36	0.391	0.938	6.14	8.24
27-30	8.9	25.8	66.7	34.5	13.3	2.59	0.421	0.940	4.74	7.50
32-37	12.3	34.3	84.0	35.8	14.6	2.45	0.400	0.946	5.66	8.04
42-48	14.2	38.4	91.2	37.0	15.6	2.37	0.392	0.938	6.36	8.62
48-53	13.8	37.8	90.4	36.5	15.3	2.38	0.393	0.954	6.27	8.42
56-62	15.4	41.9	98.7	36.8	15.6	2.36	0.391	0.934	6.35	8.55
62-67	12.7	35.3	86.3	36.0	14.7	2.45	0.400	0.948	5.76	8.18
67-72	12.1	43.3	83.3	28.0	14.5	1.93	0.321	0.930	5.11	4.94
75-81	15.6	43.0	110	36.2	14.2	2.55	0.409	0.942	5.37	8.25
84-90	15.8	43.3	110	36.5	14.3	2.55	0.409	0.946	5.51	8.42
90-96	15.0	41.7	97.2	36.0	15.4	2.34	0.388	0.949	6.26	8.17

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INVESTIGATIONS CONCERNING THE TRANSPORT OF HEAT THROUGH A SOLID SHEET OF ICE IN PERIODS OF FROST AND THAW

H. VAN WIJNGAARDEN

Senior Engineer of the Rijkswaterstaat Arnhem (Holland)

SUMMARY

Within the scope of investigations of factors having influence on the appearance and the disappearance of ice on the branches of the River Rhine in the Netherlands, it is important to pay attention to the heat-transport through a solid ice-sheet.

Investigations up to now referred to the heat-flow through a solid ice-sheet above still water. At the first suitable opportunity, it is the intention to extend the investigations and to do observations of the heat-flow through a solid ice-sheet above running water.

So called heat-flow meters were used for measurements. Such a heat-flowmeter is disc-shaped. The thickness is small. The disc is made of polyvinyl chloride, containing about 45 thermo-elements/cm² in series connection. During frost some heat-flowmeters were frozen in in a solid ice-sheet. Also some were fixed against the underside of the ice-sheet and some were lowered to the bottom.

A part of the ice-sheet in which one of the heat-flowmeters was frozen in, was covered with coal-dust; the other part was kept free.

Investigations were done to determine to what extent the heat transport through a solid ice-sheet was influenced in the time of frost and thaw by the presence of the coal-dust on the surface of the ice-sheet.

Some data could be collected concerning the influence of direct radiation of the sun on the heat flow through the heat-flowmeters in the uncovered part of the ice-sheet and in the part covered with coal dust.

While the measurements of heat transport were going on, measurements of temperature were done at the surface, within and under the solid ice-sheet with the help of thermo-couples.

The observations concerning heat transport and temperature were registered with the help of a so called Brown-recorder.

1. GENERAL

The setting in of thaw is mostly attended with a more or less important increase of the river-discharge, causing the danger of inundations in case the ice is not removed in time.

Generally spoken the presence of solid ice on the Netherlands River Rhine branches fig. 1 does not raise problems before the thaw sets in. Up to the present, the active combating of ice concentrates on clearing away the solid ice on the rivers Waal and Lek with the help of ice-breakers. Mostly these ice-breakers are put into operation before the end of a frost-period, in order to have the river ice-free as far as possible when thaw sets in. Ice breaking is only useful on the rivers Waal and Lek, because by these rivers the broken ice is discharged to the North Sea which is ice-free. This is different for the ice on the river Yssel. This Rhine-branch discharges into the Ysselmeer, which is a fresh water-basin. The Ysselmeer is covered with a solid ice-sheet very soon after the setting in of frost and remains so till long after the setting in of thaw. Under these circumstances it is of no use to break the ice on the Yssel, as the disjointed ice-floes cannot be discharged.

The execution of riverworks in the Delta-area of the Rhine, namely the canalisation of Lek and Lower Rhine and the damming-up of the sea-arms, will bring along changes concerning the appearance and disappearance of solid ice on the Rhine branches. Investigations are going on in order to determine these influences as accurately as possible.

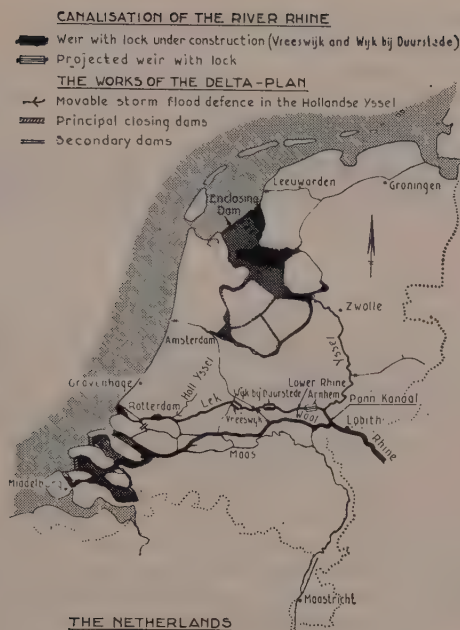


Fig. 1 — General view with schemes of Rhine-canalisation and delta-plan

Also regarding the combating of ice the question is being examined whether it is possible and necessary to look for new ways. With a view to this, investigations have been started into the process by which a solid ice-sheet is formed in periods of frost and disappears after the setting in of thaw, as well as into methods of influencing these processes by covering the ice with a layer of coal-dust and such like.

The experiments carried out up to now concern the measurement of the heat-transport through a solid ice-sheet under different circumstances. In combination, measurements of temperature and thickness of ice were done.

In the following paragraphs an exposition will be given concerning the instruments employed and the results obtained. Also attention will be given to the drawbacks of the method of investigation.

2. PURPOSE OF THE MEASUREMENTS OF HEAT-TRANSPORT AND TEMPERATURE

The purpose of the measurements of heat-transport and temperature carried out so far, can be described as follows:

1) The testing of the usefulness of the applied instruments for such measurements as well as for measurements on a more extensive scale.

2) The obtaining of a preliminary insight into the degree to which different factors can influence the heat-transport through an ice-sheet. Factors, demanding attention in the first place, are:

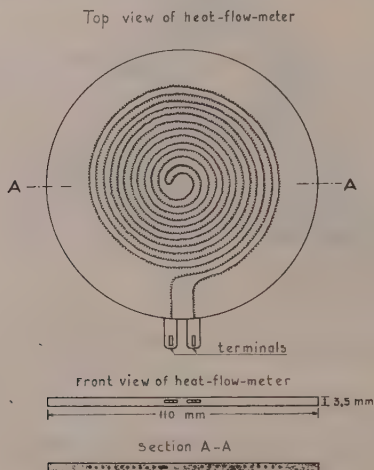
a) the influence of the temperature of the air.

b) the influence of day and night (sun-radiation and nightly radiation of the ice-surface).

- c) the influence of wind, rain-showers etc.
- d) the influence of covering the ice-sheet with coal-dust.

3. DESCRIPTION OF THE APPLIED INSTRUMENTS

The measurements of heat-transport were carried out with heat-flowmeters, developed by the "Technisch Physische Dienst T.N.O." at Delft, Holland (Fig. 2).



A heat-flow of about 4 Kcal/h.m² causes a tension of 1 mV at the terminals of the meter

Data of heat-flowmeter:

Internal resistance: about 500 Ω

Coefficient of heat-conduction: 0.25 Kcal/°C.h.m

Number of thermo-elements: about 45 per cm²

The meter is imbedded in polyvinylchloride

Fig. 2 — Heat-flow-meter

These meters are based on the principle of thermo-electric couples. If a heat-flow passes through an area F of a material with a coefficient of heat-conduction λ , a difference of temperature Q will exist between the boundaries of a layer with a thickness Δt .

In the stationary situation the relation between these data is given by:

$$Q = \lambda \cdot F \cdot \frac{\Delta t}{d}$$

in which:

Q = heat-flow in kcal/h.

λ coefficient of heat-conduction in kcal/m°C h.

F = area in m².

Δt = difference of temperature in °C.

d = thickness of layer in m.

In principle the heat-flow Q can be calculated from a measured difference of temperature over a layer with thickness d of a material with a known coefficient of

heat-conduction. The heat-flowmeter developed by T.N.O. measures the difference of temperature by means of thermo-electric couples. In order to be able to measure as accurately as possible and yet have a low heat-resistance of the measuring-disc, a large number of thermocells per cm^2 is fitted into the disc, while keeping its thickness as small as possible. By means of a special process about 45 thermocells per cm^2 have been fitted and have been electrically connected in series. The disc has a diameter of 11 cm, a thickness of 0,3 cm and is manufactured from polyvinyl chloride. After the measurement of the difference of temperature Δt , the heat-flow Q can be calculated with the help of the formula

$$Q = \lambda \cdot F \cdot \frac{\Delta t}{d}$$

in case the constants λ , d en F are known.

The electric tension at the terminals of the heat-flowmeter have been measured and recorded with the help of a potentiometric recorder (Honeywell-Brown).

Simultaneously with the measurements of the heat-flow through a solid ice-sheet, measurements of temperature were carried out, in a number of points at the surface, within and under the solid ice-sheet.

The temperature of the air was determined with a normal thermometer. The temperature at the surface, within and under the solid ice-sheet was measured with the help of thermocells. In case two conductors are connected in two joints, a difference of temperature between the joints will cause a thermo-electric tension, within a certain reach directly proportional to the difference of temperature. For the iron-constantane couples used in these experiments, the thermo-electric tension amounts to 49μ V per $^\circ\text{C}$ of difference of temperature, in the reach from 0 to 20°C .

When the thermo-electric tension occurring between the joints of a thermo-electric couple is measured accurately, it is possible to calculate the difference of temperature between the joints with the help of the constant mentioned. In order to be able to measure not only differences of temperature but also the correct value of the temperature, it is necessary to have an accurately known calibration-temperature available. The temperature of melting ice being 0°C , in case of correct handling, was found suitable for the purpose. During the thermo-electric measurements of temperature, the temperature of one joint was made 0°C by placing it in a specially constructed reservoir filled with melting ice, whilst the second joint was placed at the point where the temperature was to be determined. The tension, generated in the thermocells again were measured and recorded by a potentiometric recorder.

At first, a single-point continuous trace-recorder was used to measure and record the thermo-electric tensions, generated in the heat-flowmeters and in the thermo-electric couples. As, however, different heat-flowmeters had to be observed, a switch-clock was inserted. This switch-clock was adjusted in such a way, that the results of the measurements of each of the heat-flowmeters was recorded during 5 minutes, after which the clock switched over automatically to the next instrument. For the measurements carried out during the winter 1959-1960 a twelve-point recorder was used, making possible twelve readings almost simultaneously.

4. GENERAL PLAN OF THE MEASUREMENTS

The measurements carried out so far took place during the winters 1958-1959 and 1959-1960. As during these winters no periods of frost occurred, sufficiently severe for a solid ice-sheet to be formed on the Netherlands Rhine branches, observations were done on the ice-sheet of a frozen pond in the vicinity of Arnhem. Under the ice-sheet of the pond the water was almost stagnant.

In this respect, therefore, circumstances differed considerably from those of a solid ice-sheet in a river under which water is running. This was no objection for the investigation under consideration, because it was of importance first of all to test the usefulness of the instruments and of the method of measurements and secondly to study the behaviour of the solid ice-sheet under different circumstances.

4.1. Experiments carried out during the winter 1958-1959

On the solid ice-sheet of the above-mentioned pond two fields were indicated where the experiments were to be carried out. In both fields a heat-flowmeter was placed. In the surface layer of the ice-sheet holes were cut of about 1 cm deep. After placing the heat-flowmeters in these holes, the remaining space was filled with water. In this way, after this water was frozen, the heat-flowmeters had been completely inserted into the ice-sheet. One of the fields was covered with coal-dust (650 gr/m^2), the other was not.

Measurements were taken of the heat-flow through the heat-flowmeters, the temperature of the air in the shade, the temperature on the surface of the ice-sheet, the temperature of the water at a depth of 16 cm and, lastly, the thickness of the ice. The measurements of temperature were taken with common thermometers. The measurements were carried out at the end of a short period of frost at temperatures varying from just under to just above the freezing-point.

When at the end of the period of frost the heat-flowmeter in the field covered with coal-dust came loose, it was fixed to the underside of the uncovered field. Lastly, at the end of the investigations, a third heat-flowmeter was put on the bottom of the pond under the uncovered field. The depth of the pond at that place was about 0,80 m.

4.2. Experiments, carried out during the winter 1959-1960

Again two experimental fields were arranged on the solid ice-sheet of the formerly mentioned pond; one field was covered with coal-dust (535 gr/m^2) whereas the other field was kept clean.

In each of the experimental fields three heat-flowmeters were placed almost vertically under each other: one was inserted again under the surface of the ice-sheet, a second one was fixed against the underside of the ice-sheet, whereas the third one was placed on the bottom of the pond.

Furthermore, in both fields the temperature was measured in some places in a vertical by means of thermo-electric couples, whereas the temperature of the air in the shade was measured with a common thermometer. Also, the thickness of the ice was measured regularly.

5. RESULTS OF THE MEASUREMENTS

5.1. Measurements during winter 1958-1959

Fig. 3 shows the results of the measurements of temperature and of heat-flow.

During the 7-day period of observation the temperature of the air varied from $+8^\circ\text{C}$ to -3°C . The highest temperatures of the day occurred at about 16 hours in the afternoon. Of the air temperatures during the night only a few are known. In connection with the course of the air-temperatures by day they are, however, sufficient to get on impression of the course of temperatures by night.

In the same period the temperatures at the ice-surface varied within much narrower limits, namely between $+2^\circ\text{C}$ and $-1/2^\circ\text{C}$. Still, in this course of temperatures a direct correlation can be distinguished with the air-temperatures.

The correlation with the air-temperatures is much closer when considering the

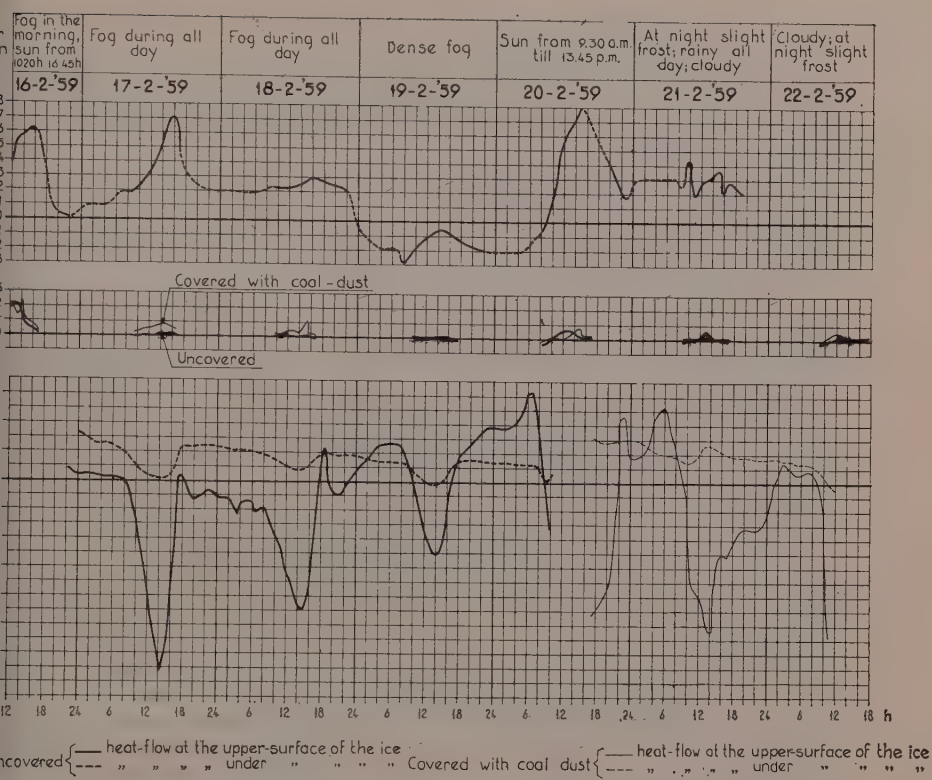


Fig. 3 — Results of temperature- and heat-flow measurements carried out in 1959.

course of the heat-flow, measured just underneath the surface of the uncovered experimental field.

In explanation of the observed course, the following remarks can be made. A downward heat-flow at the surface of the ice-sheet points to a transfer of heat from the air to the ice-sheet. This is only possible when a positive difference of temperature exists between the air and the ice-sheet. During the period that the air-temperature is considerably above 0°C , the downward heat-flow will cause the ice-sheet to melt at the surface.

It is interesting, first of all, to compare the maxima of the downward heat-flow on the 17th, 18th and 19th of February 1959 with the simultaneously observed temperatures of air and ice.

Date	max. downward heat-flow in $\text{kcal/m}^2 \text{ h}$	corresponding temperature of air in $^{\circ}\text{C}$	corresponding temperature of ice in $^{\circ}\text{C}$
17.2.59	65	+ 6°	+ $0,2^{\circ}$
18.2.59	45	+ $2\frac{1}{2}^{\circ}$	+ $0,1^{\circ}$
19.2.59	25	- $\frac{1}{2}^{\circ}$	- $0,1^{\circ}$

During the night the downward heat-flow at the surface of the ice-sheet decreases and at a certain moment changes its direction.

In the night of 17th to 18th february 1959 the heat-flow is directed upward during approximately one hour only; in the night and early morning of the 19th february this is the case during approximately 10 hours and in the night of 19th to 20th february even during 15 hours.

This change in direction and magnitude of the heat-flow during the night is in agreement with the course of temperatures.

An upward heat-flow at the surface of the ice-sheet points to a situation, in which the air-temperature is lower than the ice-temperature. In case the temperature of the air is under 0°C , melting water on the ice-surface will freeze.

Another heat-flowmeter was fixed underneath the uncovered field. From the observations at this place it could be established that the heat-flow was generally directed upward. Whether such an upward heat-flow causes an increase of the thickness of the ice-sheet depends on the temperature of the ice at the under-surface. No measurements were done in this connection, during the winter 1958-1959.

The influence of the change of air-temperature can also be observed in the course of the upward heat-flow at the under-surface of the ice-sheet. A rise of the air-temperature causes the upward heat-flow to decrease.

The general picture of the heat-flow at the upper-and at the under-surface of an ice-sheet during frost and durin thaw is schematically given in fig. 4.

The conclusions regarding the corresponding weathertype that can be drawn from the course of the heat-flow, generally appear to be correct. Only on the 19th of february the weather differs from what could be anticipated on account of the observed heat-flow.

In connection with the observations concerning the heat-flow through the uncovered experimental field, heat-flow measurements were also done in a field covered with coal-dust. Comparison of the heat-flow through the covered and through the uncovered field is hardly possible, because the measurements were not carried out simultaneously. The picture of the heat-flow through the field covered with coal-dust agrees in general lines with that observed in the uncovered field.

The measured heat-flow at the surface is strongly related to the air-temperature; when at night the air-temperature drops beneath 0°C , the heat-flow is directed upward; by day, at temperatures above the freezing-point the heat-flow is directed downward. The heat-flowmeter at the under-surface of the ice-sheet generally shows an upward heat-flow in this case too.

In order to gain an insight in the quantitative significance of the measured heat-flow, a comparison was made between the measured change in the thickness of ice and the quantities of heat added to the ice-sheet.

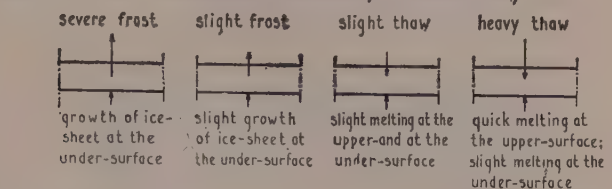
On the 17th, 18th and 19th february 1959 the thickness of the ice was measured, while the quantity of heat supplied during these days is known.

The area between the lines indicating the measured course of the heat-flow at the under- and at the upper-surface, is a measure for the quantity of heat supplied to the ice-sheet.

In view of the difficulty of determining the thickness of the ice with sufficient accuracy, the measured changes in this value agree rather well in order of magnitude with what could be anticipated on account of the data concerning the supplied quantities of heat.

In addition, during the period of observation in 1959 some data have been collected concerning the relation between the measured heat-flow through the heat-flowmeters at the surface of the ice-sheet and the intensity of the sun-radiation. In fig. 5 the results have been compiled.

General picture of the expected heat-flow at the upper-and at the under-surface of an ice-sheet during frost and during-thaw.



THE OBSERVED HEAT-FLOW SHOWS THE FOLLOWING PICTURE.

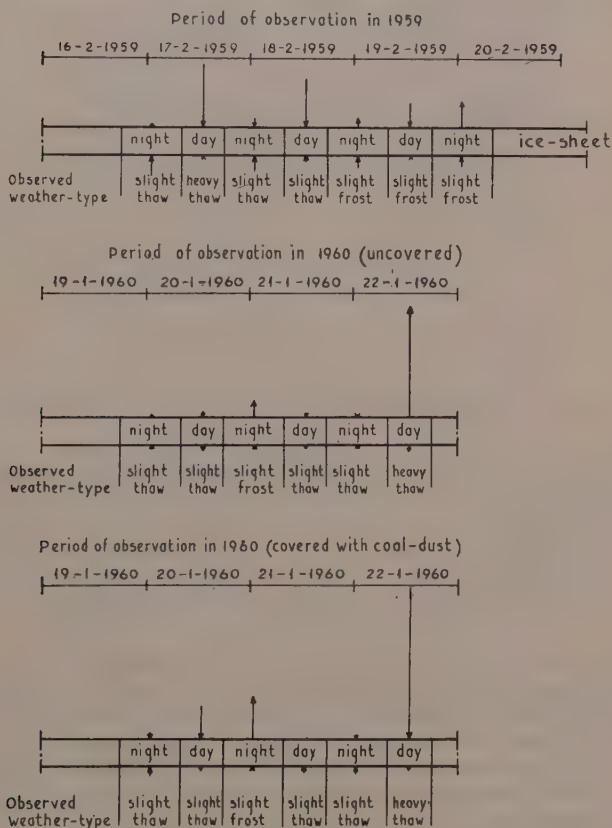


Fig. 4 — Heat-flow-through an ice sheet.

The intensity of sunlight on the ice-sheet was taken to be a measure of the intensity of the sun-radiation.

It appears that for the uncovered field a close relation exists between the intensity of light and the heat-flow. The fluctuations of light-intensity are, with a slight retardation, found again in the course of the observed heat-flow.

For the field covered with coal-dust too a correlation, though in more general lines, can be established between light-intensity and heat-flow. In this case however,

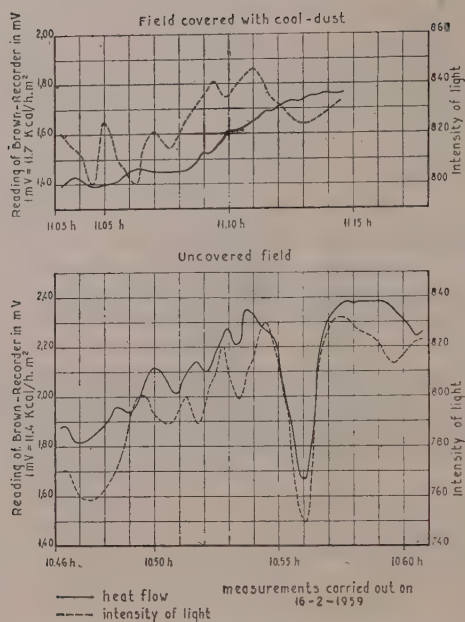


Fig. 5 — Relation between intensity of light and heat flow

short fluctuations of the light-intensity do not appear in the observed heat-flow. This may be caused by heat-accumulation of the coal-dust. By some simple experiments it could be established, that the reactions of the heat-flowmeter are caused only by fluctuations of the sun-radiation and not by the corresponding changes of light-intensity. For these experiments two sources of light were utilized, issuing different quantities of heat-radiation. One was a normal electric bulb, the other a fluorescent lamp.

First, the electric bulb was placed opposite the heat-flowmeter and measurements were done of the light-intensity on the meter and of the indication of the galvanometer connected to it. Then the distance of the fluorescent lamp to the heat-flowmeter was chosen in such a way, that the same light-intensity on the meter was measured. The difference of the readings of the galvanometer in the first and in the second instance was of the order to be expected from the known data concerning heat-radiation of the two light-sources.

Lastly it is interesting to note that rain- and hail-showers had a direct influence on the course of the heat-flow, through the meter at the surface of the ice-sheet (fig. 6).

5.2. Measurements during winter 1959-1960

In fig. 7 a number of measuring-results have been compiled.

The period of the measurements included only one night during which the temperature dropped below the freezing-point for a reasonably long time. During the hours that the temperature of the air dropped beneath 0°C , also a dropping of the temperature at the surface of the ice-sheet is to be observed. The temperature at the bottom as well seems to show a direct connection with the air-temperature.

THE HEAT-FLOWMETER, IS FROZEN IN
AT THE SURFACE OF THE ICE-SHEET.

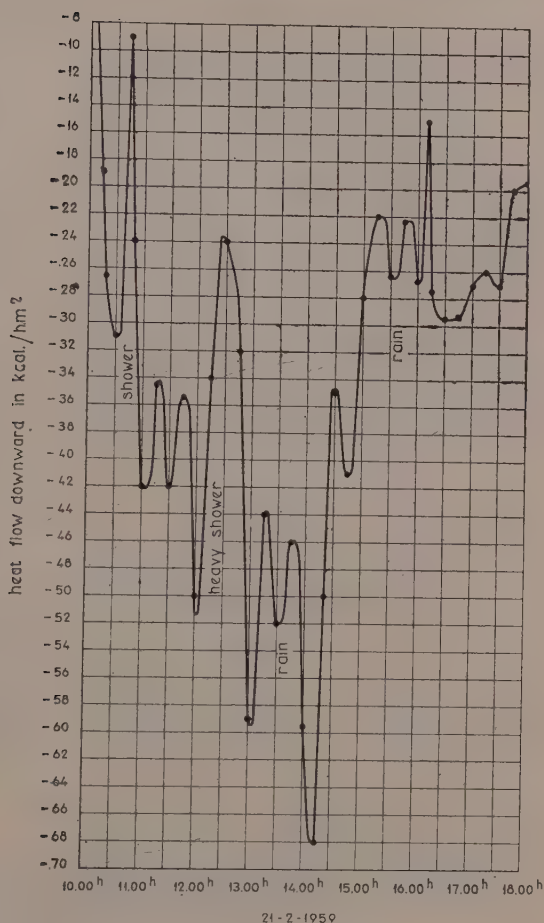


Fig. 6 — Influence of rainshower on heat-flow.

Concerning the observations with heat-flowmeters in both fields, the following remarks can be made:

a) The heat-flowmeters at the surface of the ice-sheet show a direct and sharp reaction to fluctuations of the air-temperature.

In case the air-temperature is below freezing-point, an upward heat-flow is observed; at temperatures above 0°C this heat-flow is directed downward.

b) The heat-flowmeters at the under-surface of the ice-sheet generally indicated, during the period of observation, a small upward heat-flow. Nevertheless here too a correlation with the changes of air-temperature can be observed. The fluctuations of the air-temperature are only weakly reflected by the course of the heat-flow.

c) The measured heat-flow at the bottom is small and directed upward. The

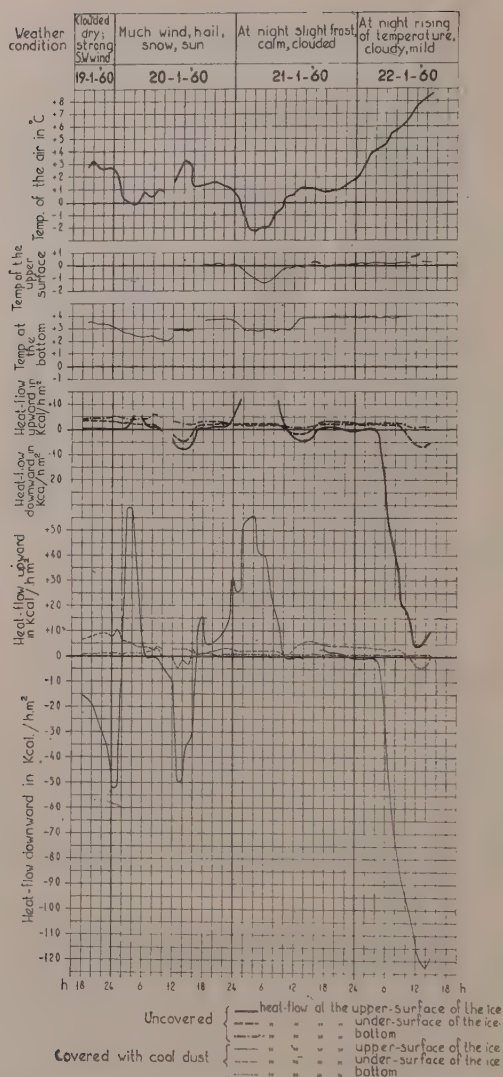


Fig. 7 — Results of temperature- and heat-flow measurements carried out in 1960.

fluctuations are small and do not show much correlation with the variations of the air-temperature.

d) In fig. 4 again a schematic picture is given of direction and magnitude of the measured heat-flow at the upper- and at the under-surface of the covered field and of the corresponding type of weather.

e) The number of data available is still too small to make a quantitative comparison possible between the heat-flow through the uncovered field and through the covered field.

Nevertheless it can be noted, that for the given weather-conditions the fluctuations as well as the absolute magnitude of the heat-flow through the uncovered field are smaller than those through the covered field.

More extensive observations at different temperatures and under different weather-conditions are required, in order to be able to give a well-founded opinion.

Disturbance of the heat-flow in a solid ice-sheet by the presence of a heat-flowmeter

In future efforts will be made to utilize the heat-flow measurements in a more quantitative sense. For this purpose, one of the questions to be answered is: to what degree does a heat-flowmeter disturb the heat-flow through a solid ice-sheet? A disturbance must be anticipated on account of the fact that the heat-flowmeter has a coefficient of heat-conduction different from that of ice. An attempt will be made to find the relation between the heat-flow through a homogeneous ice-sheet and the heat-flow to be observed through a frozen-in heat-flowmeter.

This investigation will be only concerned with the transfer of heat by conduction. The possible influence of heat-radiation will be attended to in the next paragraph.

The calculation of the heat-flow through a disc with an area F and a thickness d is based on the formula:

$$Q = \lambda \cdot F \cdot \frac{\Delta t}{d}$$

in which:

Q = heat-flow in kcal/h

λ = coefficient of heat-conduction in kcal/h m°C

F = area in m²

Δt = difference of temperature in °C

d = thickness of ice-sheet in m.

For a homogeneous ice-sheet (without heat-flowmeter) with a thickness d , with a temperature at the upper-surface of t_1 °C and at the under-surface of t_4 °C (Fig. 8), applies:

$$Q = \lambda_i \cdot F \left(\frac{t_1 - t_4}{d} \right)$$

λ_i = coefficient of heat-conduction of the ice in kcal/h m°C.

Supposing a heat-flowmeter of infinite area is frozen in an ice-sheet, the following relation is found (fig. 8):

$$t_1 - t_4 = (t_1 - t_2) + (t_2 - t_3) + (t_3 - t_4)$$

$$= \frac{Q_w d_1}{\lambda_i F} + \frac{Q_w d_2}{\lambda_w F} + \frac{Q_w d_3}{\lambda_i F}$$

$$= \frac{Q_w}{\lambda_i F} \left(d_1 + \frac{\lambda_i}{\lambda_w} d_2 + d_3 \right)$$

$$Q_w = \frac{\lambda_i F (t_1 - t_4)}{d_1 + \frac{\lambda_i}{\lambda_w} d_2 + d_3}$$

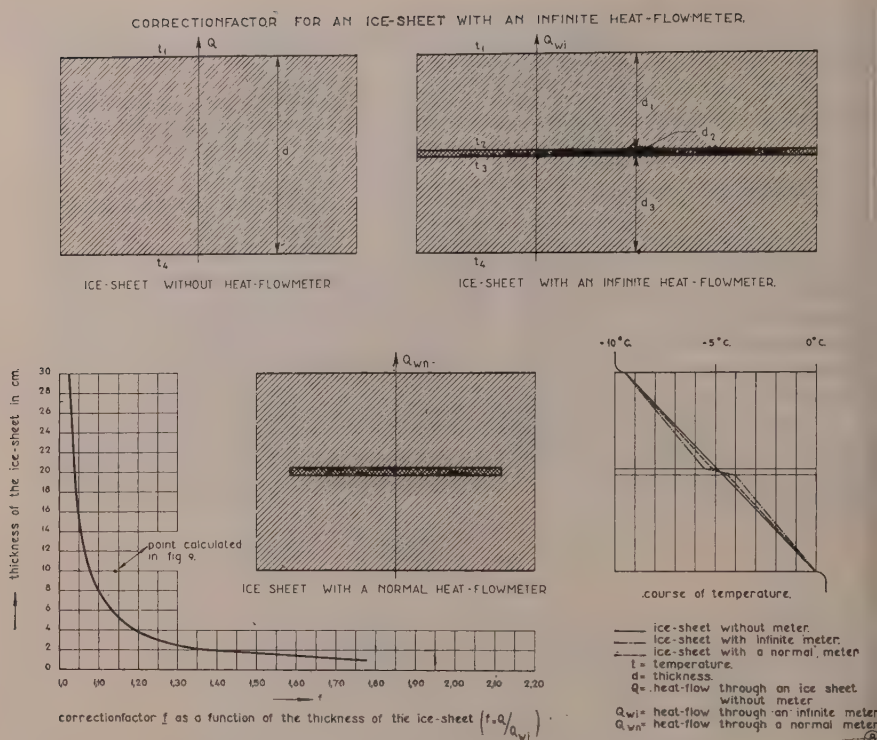


Fig. 8 — Heat-flow trough and course of temperature in an ice-sheet.

in which:

Q_w = heat-flow through heat-flowmeter and ice-sheet in kcal/h.

λ_w = coefficient of heat-conduction of the meter.

In order to reduce the measured heat-flow Q_w to the heat-flow through a homogeneous ice-sheet without a heat-flowmeter, a correction-factor f is introduced

$$Q = fQ_w$$

$$f = \frac{Q}{Q_w} = \frac{d_1 + \frac{\lambda_i}{\lambda_w} d_2 + d_3}{d}$$

in which:

$\lambda_i = 0.8 \text{ kcal/}^\circ\text{C m h}$

$\lambda_w = 0.25 \text{ kcal/}^\circ\text{C m h}$

$d_2 = 0.0035 \text{ m}$

These values inserted in the formula give:

$$f = 1 + \frac{0.0077}{d}$$

In fig. 8 this correction-factor is represented graphically. From the figure it will be noted that the heat-flow through a homogeneous ice-sheet, thick 10 cm, (without a heat-flowmeter) will exceed the heat-flow measured with the meter by $7\frac{1}{2}\%$.

In addition fig. 8 gives a picture of the course of the temperature in an ice-sheet without a heat-flowmeter, in an ice-sheet with an infinite heat-flowmeter and in an ice-sheet in which a heat-flowmeter of normal dimensions is placed.

The course of temperatures through an ice-sheet, in which a normal heat-flowmeter is placed, will show values between the indicated course for the other two situations, in a cross-section through the centre of this meter. This means, that the temperature gradient in the centre of the normal heat-flowmeter and considered over the thickness of the meter, will be smaller than the corresponding gradient in the infinite meter. The heat-flow through a heat-flowmeter of normal dimensions will consequently under otherwise equal circumstances be smaller than the heat-flow through an infinite meter.

The correction-factor to be applied to the observations with a normal heat-flowmeter will have to be larger than the one indicated in fig. 8 for an infinite heat-flowmeter.

In case the picture of the heat-flow in an ice-sheet is known in and around the place of a normal heat-flowmeter, the correction-factor to be utilized can be determined.

As an example this picture, offered by the heat-flow in a certain case was determined by iteration. The result is shown in fig. 9.

Two systems of lines perpendicular to each other were drawn. One of the systems consists of lines of equal temperature, the other consists of so-called "stream" lines of the heat-flow.

A condition is, that the heat-flow should be the same in the different part of a streamlane. This was checked, and with the help of the deviations found, the preliminary system was revised. By repeating this procedure several times, at last a satisfactory system was obtained. The corresponding correction-factor amounted to: $f_1 = 1,145$.

For the same case, but with an ice-sheet with an infinite heat-flowmeter, was found: $f = 1,075$.

The correction-factor for the observations with a normal heat-flowmeter will, as was the case with an infinite meter, depend on the thickness of the ice-sheet.

Moreover it seems likely, that this correction-factor depends on the place of the heat-flowmeter in the ice-sheet.

In order to know the correction-factor for different thickness of the ice-sheet, heat-flow-pictures must be determined by iteration by a method corresponding to the one set forth above.

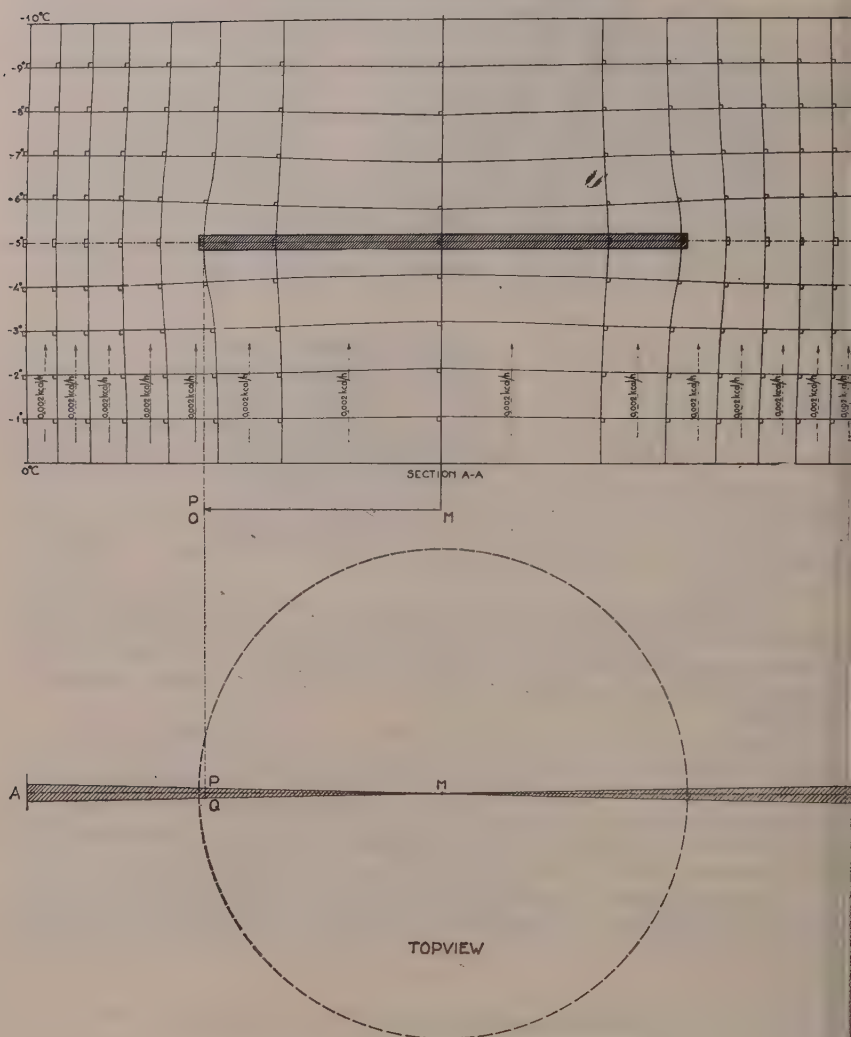
Influence of heat-radiation on the heat-flowmeter during measurements concerning the heat-flow through an ice-sheet.

In fig. 5 the results are shown of the relation between the intensity of direct or indirect sunlight falling on the heat-flowmeter, and the observed heat-flow.

As set forth before, the measured intensity of light can be regarded as a measure for the sun-radiation falling on the meter. Clearly, the heat-flow through the meter at the upper-surface of the ice-sheet is influenced directly by the sun-radiation.

The heat-flowmeter does not only give an insight into the heat-flow through the ice-sheet caused by the transfer of heat by conduction, but also an impression of the quantity of heat from the sun reaching by radiation the spot where the heat-flowmeter is placed. The sun-radiation is partly caught at the upper-surface of the heat-flowmeter, because the meter is less pervious than the surrounding ice. In this way

HEAT-FLOW LANES AND LINES OF EQUAL TEMPERATURE FOR AN
ICE-SHEET WITH A HEAT-FLOWMETER FROZEN IN



The calculation was carried out for the shaded sector

Data: coefficient of heat-conduction for ice 0.5 kcal/cm^2
coefficient of heat-conduction for heat-flowmeter 0.25 kcal/cm^2

Result: through the sector MPQ of the
heat-flowmeter flows 0.004 kcal/h ;
through an equal sector, without
heat-flowmeter, flows 0.00458 kcal/h .

$$f = \frac{Q}{Q_0} = \frac{0.00458}{0.004} = 1.145$$

Fig. 9

the temperature at the upper-surface rises, causing the heat-flow to increase. Thus, the difference between the heat-flow measured through the heat-flowmeters at the upper- and at the under-surface of the ice-sheet can partly be explained by the difference of radiation -heat penetrating to each of the meters.

In view of all this it must be stated, that the measured heat-flow certainly does not always give an exact picture of the heat-transport through the ice-sheet.

By comparison of the measured heat-flow at the upper- and at the under-surface of the ice-sheet it is, however, possible to obtain an insight into the total quantity of heat, supplied to or withdrawn from the ice-sheet by conduction and radiation combined.

HEAT FLOW IN ICE SHEETS AND ICE CYLINDERS ¹

E.R. POUNDER

RÉSUMÉ

L'équation de l'écoulement de chaleur est résolue pour deux modèles simples. Dans le premier cas, une couche de glace, avec un gradient uniforme de température (de la température de l'air à celle de l'eau), est soumise à une réduction de la température de l'air en forme de fonction en escalier. On montre qu'il y aura un retard de l'ordre de jours avant que la température réduite de l'air cause une augmentation dans le taux de gel de la glace. Dans le deuxième cas, on fait mouvoir un cylindre de glace ayant une température uniforme dans une température nouvelle de l'air ambiant. Le temps pris par le cylindre pour atteindre la nouvelle température est calculé. Les deux analyses négligent l'effet de la couche limite de l'air à la surface de la glace et les résultats expérimentaux présentés montrent l'influence importante de cette couche.

SUMMARY

The heat flow equation is solved for two simple models. In the first, an ice sheet with a uniform temperature gradient (from the air temperature to the water temperature) is subjected to a step function drop in air temperature. It is shown that there will be a delay of the order of days before the lower air temperature causes an increase in the freezing rate of the ice. In the second model, a cylinder of ice at constant temperature is moved to a new ambient air temperature. The time for the cylinder to reach the new temperature is calculated. Both analyses neglect the effect of the boundary layer at an air-ice interface, and experimental results showing the major influence of this layer are reported.

1. INTRODUCTION

In a previous paper (Pounder and Stalinski 1960), observations made near Resolute Bay were reported in which it was shown that several days elapsed between a sharp drop in the air temperature and a subsequent increase in the rate of formation of new ice on the bottom of the ice cover. The new ice showed the higher salinity characteristic of more rapid freezing of salt water. This paper attempts a theoretical analysis to obtain the order of magnitude of this delay.

A similar problem in heat conduction occurs when ice properties are studied from cores extracted from an ice cover. The air temperature is usually colder than that of most of the ice cover and it may be of importance to know how rapidly the temperature of the core changes. The infinite cylinder is one of the standard problems in heat conduction (see Bowman (1958) pp. 37-39, for example), provided the effect of the boundary layer is neglected. This is a gross over-simplification of the physical phenomena involved.

2. TRANSMISSION OF A TEMPERATURE STEP THROUGH AN ICE SHEET

To evaluate the order of magnitude of the transmission delay of a temperature step, the simple model of Fig. 1 is considered. It is assumed that the air temperature has remained constant at θ_0 for a long time and that the thermal conductivity of

(¹) Une contribution du projet de la recherche sur la glace, Département de Physique, Université McGill, Montréal, Canada.

(¹) Contribution from the Ice Research Project, Department of Physics, McGill University, Montreal, Canada.

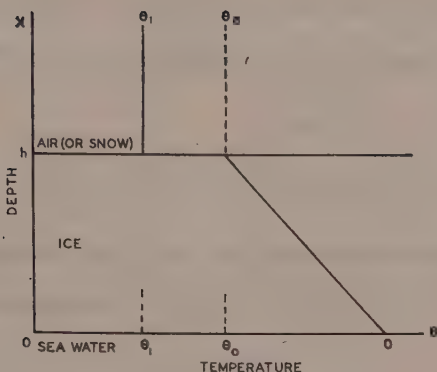


Fig. 1 — A model for calculating the effect on an ice cover of a step function change in the air temperature.

the ice is a constant, so that there will be a linear temperature gradient through the ice. For convenience of calculation a temperature scale in centigrade degrees, but with a zero at the freezing point of sea water, is chosen. If temperature on this scale is called $^{\circ}\text{R}$,

$$\theta^{\circ}\text{C} = \theta^{\circ}\text{R} - 1.9. \quad (1)$$

At time $t = 0$ a step change in air temperature, to θ_1 , is assumed to occur. It is further assumed that the diffusivity $K = \frac{k}{sd}$ is a constant. Here k , s , and d are the thermal conductivity, specific heat, and density of the ice. In sea ice, both k and s actually change rapidly with depth in the ice sheet, because of variations in temperature, salinity, and crystal structure. The change in the thickness h of the ice sheet during the time interval considered is neglected.

The equation of heat flow is

$$\frac{\partial \theta}{\partial t} = K \frac{\partial^2 \theta}{\partial x^2}$$

with boundary conditions

$$\theta(h, t) = \theta_1, \theta(0, t) = 0, \theta(x, 0) = \theta_0 \frac{x}{h},$$

where h is measured vertically upwards from the water-ice interface.

Change the dependent variable to $\varnothing(x, t)$ where

$$\varnothing(x, t) = \theta(x, t) - \theta_1 \frac{x}{h}. \quad (2)$$

The function \varnothing satisfies the equation

$$\frac{\partial \varnothing}{\partial t} = K \frac{\partial^2 \varnothing}{\partial x^2}$$

with boundary conditions

$$\varnothing(h, t) = 0, \varnothing(0, t) = 0, \varnothing(x, 0) = (\theta_0 - \theta_1) \frac{x}{h}.$$

Put $\varnothing(x, t) = X(x) T(t)$. Then

$$\frac{1}{T} \frac{dT}{dt} = \frac{K}{X} \frac{d^2 X}{dx^2} = \text{a separation constant, } -\beta^2$$

and

$$T = T_0 \exp(-\beta^2 t)$$

$$X = A \sin(\gamma x) + B \cos(\gamma x) \text{ where } \gamma = + \sqrt{\frac{\beta^2}{K}}.$$

From the boundary conditions, $X(0) = X(h) = 0$, and hence

$$B = 0, \gamma = n\pi/h, \text{ i.e. } \beta^2 = \frac{n^2 \pi^2 K}{h^2}, \text{ where } n \text{ is an integer.}$$

$$\therefore \varnothing(x, t) = \sum_n A_n \exp\left(\frac{-n^2 \pi^2 K t}{h^2}\right) \sin \frac{n \pi x}{h}. \quad (3)$$

$$\varnothing(x, 0) = (\theta_0 - \theta_1) \frac{x}{h} = \sum_n A_n \sin \frac{n \pi x}{h}.$$

The Fourier coefficient is readily evaluated as

$$A_n = (-1)^{n+1} \frac{2}{n\pi} (\theta_0 - \theta_1).$$

Substituting in (3) and combining with (2),

$$\theta(x, t) = \theta_1 \frac{x}{h} + \sum_{n=1}^{\infty} (-1)^{n+1} \frac{2}{n\pi} (\theta_0 - \theta_1) e^{-\frac{n^2 \pi^2 K t}{h^2}} \sin \frac{n \pi x}{h}. \quad (4)$$

The series will converge rapidly for large values of t . Since time enters only through the factor $\frac{t}{h^2}$, we can deduce that the time for a temperature disturbance to reach a certain fractional depth of the ice sheet increases as the square of the total thickness of the ice.

Numerical Data

The situation near Resolute approximated to this model about Nov. 1, 1958. The mean daily temperature was roughly constant from Oct. 23 to Oct. 28, then dropped rapidly by about 15 fahrenheit degrees and held roughly constant for the following week. Let θ_0 be taken as the average of the mean temperatures for Oct. 23-28 and θ_1 as the average of the mean temperatures for Oct. 29-Nov. 4. From the data in Pounder and Stalinski (1960)

$$\theta_0 = 2.4^\circ\text{F} = -16.4^\circ\text{C} = -14.5^\circ\text{R}$$

$$\theta_1 = -13.1^\circ\text{F} = -24.0^\circ\text{C} = -22.1^\circ\text{R}.$$

To calculate the diffusivity, the measured values of k (2.14×10^{-3}) and d (0.943) are used). The value of s is taken from a table given by Malmgren (1927) as $0.68 \text{ cal gm}^{-1} \text{ per } ^\circ\text{C}$ for a (mean) temperature of -10°C and a salinity of 5.1% . Hence $K = 3.34 \times 10^{-3} \text{ c.g.s. units}$. The mean thickness of the ice during this period was about 20 inches,

say $h = 50$ cm. Let us calculate the variation of temperature with time for $x' = 12.5$ cm, that is three-quarters of the way through the ice sheet from the top.

$$\begin{aligned} \theta(x',t) &= \frac{-22.1}{4} + \frac{2}{\pi} \times 7.6 \sum_{n=1}^{\infty} \frac{(-1)^{n+1}}{n} \sin \frac{n\pi}{4} \exp \left(\frac{-n^2 \pi^2 \times 3.34 \times 10^{-3} t}{50.0^2} \right) \\ &= -5.52 + 4.87 \sum_{n=1}^{\infty} \frac{(-1)^{n+1}}{n} \sin \frac{n\pi}{4} \exp(-1.32 \times 10^{-5} n^2 t). \end{aligned}$$

Evaluation of the exponential term for $t = 12$ hours, 1 day, etc. shows that after 12 hours the $n = 3$ terms are negligible and after 1 day the $n = 2$ terms are negligible. The results of calculating $\theta(x',t)$ are plotted in Fig. 2.

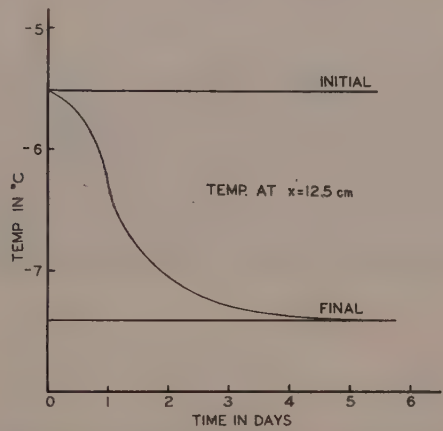


Fig. 2 — The theoretical temperature variation with time in the ice cover after a temperature step.

From the graph, the maximum increase in the freezing rate would be expected to occur after 1 to 2 days, whereas it was actually found to occur about 8 to 10 days after the sharp drop in the air temperature.

3. RELAXATION OF TEMPERATURE IN AN ICE CYLINDER

Suppose an infinite cylinder of diameter $2a$ is at a uniform initial temperature of $\theta_0^\circ\text{C}$ and is moved at zero time to a region of constant ambient air temperature $\theta_1^\circ\text{C}$. If it is assumed that the surface of the cylinder changes instantly to the new temperature, the differential equation and boundary conditions are as follows :

$$\begin{aligned} \frac{\partial \theta}{\partial t}(r,t) &= \frac{K}{r} \frac{\partial}{\partial r} \left(r \frac{\partial \theta}{\partial t} \right) \\ \theta(r,0) &= \theta_0, r \leq a \\ \theta(a,t) &= \theta_1, t \geq 0. \end{aligned} \tag{5}$$

The solution of this problem is given in Bowman (1958), p. 39, as

$$\theta(r,t) = \theta_1 + 2(\theta_0 - \theta_1) \sum_n \frac{1}{\alpha_n J_1(\alpha_n)} \left[J_0 \left(\alpha_n \frac{r}{a} \right) \right] \exp \left(- \frac{K \alpha_n^2 t}{a^2} \right), \tag{6}$$

where J_0, J_1 are Bessel functions, α_n is the n th root of $J_0(x) = 0$, and the summation is over all the positive roots of this equation.

The temperature at the centre of the cylinder is

$$\theta(0, t) = \theta_1 + 2(\theta_0 - \theta_1) \sum_n \frac{1}{\alpha_n J_1(\alpha_n)} \exp \left(- \frac{K \alpha_n^2 t}{a^2} \right)$$

since $J_0(0) = 1$. The percentage change $P_c(t)$ of the temperature at the centre is given by

$$P_c(t) = \frac{\theta(0, t) - \theta_0}{\theta_1 - \theta_0} \times 100$$

$$P_c(t) = 100 - 200 \sum_n \frac{1}{\alpha_n J_1(\alpha_n)} \exp \left(- \frac{K \alpha_n^2 t}{a^2} \right). \quad (7)$$

3.1. Numerical Values

Most of the cores studied are 7.50 cm in diameter, so that $a = 3.75$. For sea ice take $K = 3.34 \times 10^{-3}$ c.g.s. units as above. The series (7) converges rather slowly for small values of t . For $t = 1$ min., 6 terms are necessary, and $t = 15$ min, 2 terms are needed. After this interval one term is sufficient, and $P_c(t)$ is a simple function of exponential growth with a time constant $= \frac{3.75^2}{K \alpha_1^2} = 728 \text{ sec.} \approx 12 \text{ min.}$ The graph of P_c for sea-ice cores is plotted in Fig. 3. Since the transient has risen to over half its

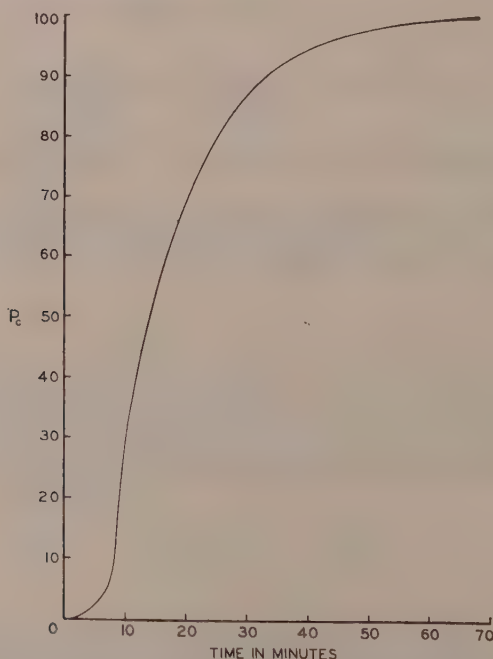


Fig. 3 — Theoretical temperature relaxation at the centre of a sea-ice core.

final value in 15 min. ($P_c = 53.6$), after an additional four time constants have elapsed $P_c > 99\%$, and the temperature change can be considered as completed in approximately one hour.

For fresh-water ice, the series must be recalculated with a different diffusivity. Taking k as 5.68×10^{-3} , s as 0.478 (the value at -15°C), and d as 0.917, all in c.g.s. units, $K = 0.0130$. This larger value of K shows that fresh-water ice should adjust more rapidly to temperature changes. Fig. 4 shows the graph of P_c for this case.

The analyses discussed above make the seriously wrong assumption that the ice surface assumes the new ambient air temperature immediately. This is contrary to the well known boundary layer effect in heat conduction, in which a thin stationary layer of air adjoins the solid surface. There is a large temperature gradient across this layer giving extra insulation against conduction, and also reducing convective cooling.

An experiment was undertaken to investigate this effect. A cylinder of fresh-water ice (7 inches long by 3 inches in diameter) was drilled axially and a thermocouple frozen in at its centre. Two ambient temperatures were chosen, -28°C in a deep freeze and -8°C in the cold room. The sample was left for a long time exposed to one of these ambient temperatures, then moved to the other, and the changes of temperature at the centre of the core were measured. Fig. 5 shows the results obtained, expressed as percentage temperature changes. The solid curve shows the changes when the sample was stored at -8°C and then transferred to the deep-freeze unit. There was no forced air circulation in the freezer and the time for 95% completion was approximately in 150 minutes, which is about 12 times as long as predicted in Fig. 4. The dashed curve in Fig. 5 represents the reverse change—from -28°C in the freezer to -8°C in the cold room. The more rapid temperature change was presumably caused by the fact

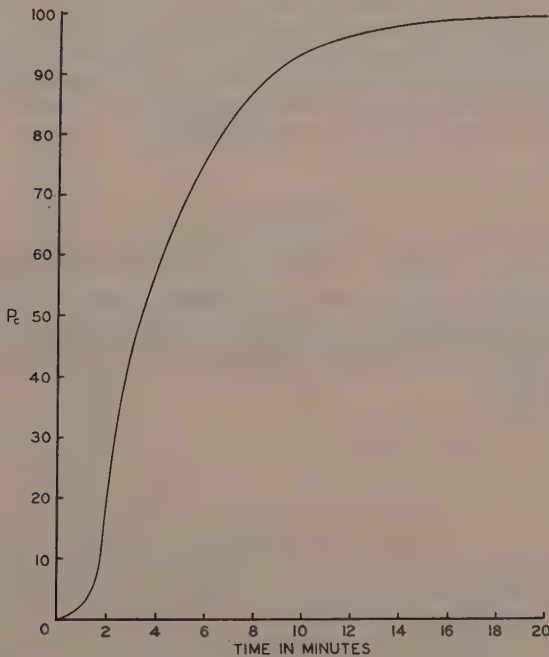


Fig. 4 — Theoretical temperature relaxation at the centre of a fresh-water ice core.

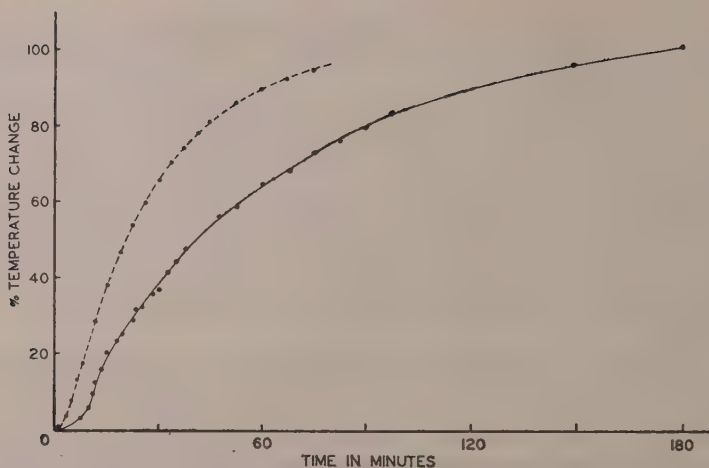


Fig. 5 — Experimental temperature relaxation curves. Fresh-water ice.

that the air in the cold room is continually circulated. The circulation is quite slow in terms of wind speed—the “wind” past the sample was less than one mile per hour. Even so, the 95% completion figure is approximately 80 minutes or about 7 times as long as the theoretical figure.

Some tentative conclusions can be drawn. If a 3-inch core of freshwater ice is removed to a new ambient temperature three hours is sufficient for it to take up the new temperature, even in the absence of any forced air circulation. The comparable maximum time for a sea-ice core is about twelve to fifteen hours.

The assistance of Dr. M. P. Langleben in making the temperature measurements on the ice core is acknowledged. This work was supported by the Defence Research Board of Canada through D. D. P. Contract GC.69-900109.

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ON THE MASS AND HEAT BUDGET OF ARCTIC SEA ICE

N. UNTERSTEINER

SUMMARY

Measurements during the drift of « US Drifting Station A » show an annual mass increase of old ice consisting of $12,5 \text{ g/cm}^2$ snow and 52 g/sm^2 bottom accretion. During the summer seasons 1957 and 1958 an amount of 19,2 and $41,4 \text{ g/cm}^2$, respectively, was lost by surface ablation. The ratio of ablation on elevated « dry » surfaces and in meltwater ponds is 1 : 2,5. The average pond area was about 40 %. Bottom ablation by heat transfer from the ocean was found to be 23 cm (July to Aug./Sept.).

Methods of measuring mass changes are described. In view of their importance as a means of checking the computed heat budget their accuracy is discussed in detail.

The heat budget is computed for a selected period during the height of the melt season. The average daily totals are : $+146 \text{ cal/cm}^2$ from net short wave radiation, -8 cal/cm^2 from net long wave radiation, $+9 \text{ cal/cm}^2$ from turbulent transfer of sensible, and -11 cal/cm^2 of latent heat. Daily surface ablation is 0,8 cm. About 90 % of it is due to the absorption of short wave radiation. Only 52 % of the total heat supply are transformed at the surface. 48 % are transmitted into the ice and mainly used to increase the brine volume. The latter acts as a reserve of latent heat during the cooling season. With a ice thickness of 300 cm its amount is approx. 3700 cal/cm^2 .

ON THE INFLUENCE OF TREES ON ACCUMULATION OF SNOW IN PINE DOMINATED FOREST IN FINLAND

SEPPÄNEN, M.
(Hydrological Office. Helsinki)

SUMMARY

The water equivalent of snow cover in pine dominated forest of different density in relation to that of open fields during the accumulation of snow cover will be discussed here. Furthermore, some computations will be presented on the influence of an individual tree in pine dominated forest on the unevenness of the snow pack.

The influence of forest density on accumulation of snow cover may approximately be derived from the snow course measurements of snow cover. Since the date February 16th, fell within the period of snow accumulation during all the years 1952-60, the average water equivalent of snow cover computed from snow course records of all stations for this date indicate the approximate amount of snow accumulated in different types of terrain on February 16. Results obtained are shown in Table 1.

TABLE 1

Average water equivalent of snow cover in Finland according to snow course measurings at all stations on February, 16, in years 1952-60 in per cent of the water equivalent of snow cover (109 mm) on open fields respectively.

Type of terrain	Water equivalent of snow cover per cent
Open fields	100
Glades	109
Pine dominated forest:	
thin	102
rather thin	100
normal	95
rather dense	93

Table 1 shows that during years 1952-60 in Finland less snow accumulated in pine dominated forest of normal density than on open fields.

A more accurate study of the influence of forest density on accumulation of snow cover may be obtained from snow stations, there a greater number of measuring sticks with centimeter scale is inserted into the ground in forest.

Fig. 1 shows the map of such a station in operation in Juuka (N. Lat. 63°6', E. Long. 29°28'). At this station there are 121 snow measuring sticks all 5 meters apart

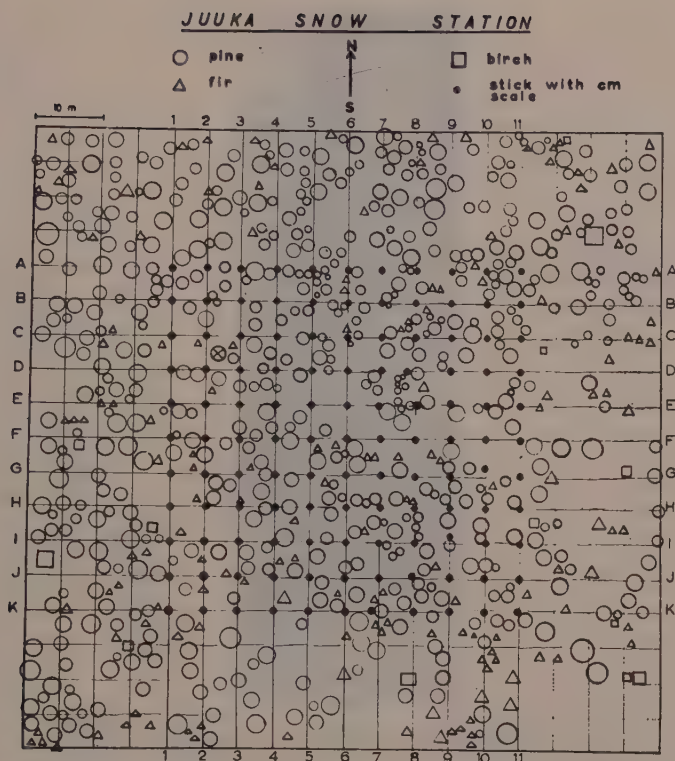


Fig. 1

arranged in a square. The calibre of trees is shown in Fig. 2, there the pine marked with a cross is also seen in Fig. 1, also marked with a cross, inside a square marked by measuring sticks C 2, C 3, D 3 and D 2. This tree is 15 meters high with a vertical profile area of 16 square meters including the trunk, which is 21 centimeters in diameter measured at the height of 1.5 meters above the ground. The vertical profile area of the pines on the measuring area varied from 1-30 square meters. The actual measurement area was 2500 square meters and the whole area mapped down measured 8100 square meters.

We will now examine how the water equivalent of snow cover in pine dominated forest at snow station in Juuka on January 15, 1960, depended upon the forest density and what sort of a significance the method of determining the density of the forest had in this respect.

We denote

w = water equivalent of snow cover (mm) at measuring stick.

w_0 = average water equivalent of snow cover of the actual measurement area (2500 square meters).

$\Delta w = w - w_0$

A = size in square meters of the area, which is used for determination of forest density.

B = vertical profile area of pine in square meters.



Fig. 2

ΣB = total of the vertical profile area of pines on an area the size of A .
 $B\Sigma_0$ = average value of expression ΣB for the whole area mapped down (8100 square meters).
 $\Delta \Sigma B = \Sigma B - \Sigma B_0$.

In this investigation forest density refers to the average share of the total vertical profile area of pines on an area the size of A falling on 1 square meter, in other words, the values of the expression $\frac{\Sigma B}{A}$. The average vertical profile area of pines falling on each square meter of the whole area mapped down was 0.86 square meters.

On January 15, 1960, the water equivalent of snow cover of the actual measuring area at the snow station in Juuka was $w_0 = 65$ mm. We will now calculate the average value of Δw in such a case, there $\frac{\Delta \Sigma B}{A} = 1$, or in other words, in such case, there the vertical profile vertical of pines close to the measuring stick on an area the size of A averages 1.86 square meters per one square meter. The result greatly depends on the size of A used in determining forest density.

Fig. 3 shows the corresponding deviation of water equivalent of snow cover Δw (mm) in relation to the deviation $\frac{\Delta \Sigma B}{A} = 1$ of forest density as a function

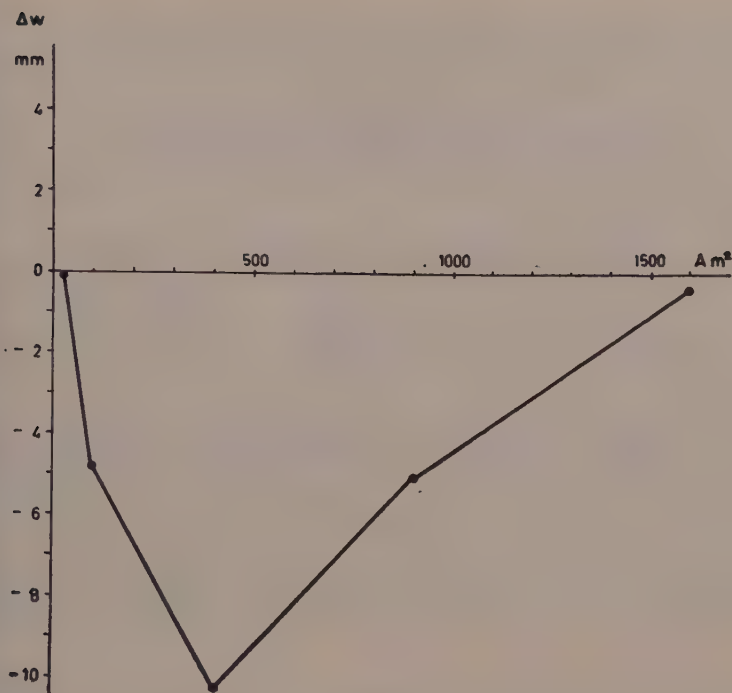


Fig. 3

of area A , which has been employed in determination of forest density at snow station in Juuka on January 15, 1960. The Fig. also shows that the values of the deviation Δw have been calculated separately for different values of A . This would indicate that if in determining forest density the area used is 400 square meters, the amount of snow in the denser part of the stand is considerably less than the average. However, on an area of 1600 square meters the variability of the forest density does not have any greater influence on the accumulation of snow.

In pine dominated forest every individual tree has an effect of its own on the unevenness of accumulation of snow. In the following a result of computation in this respect will be given.

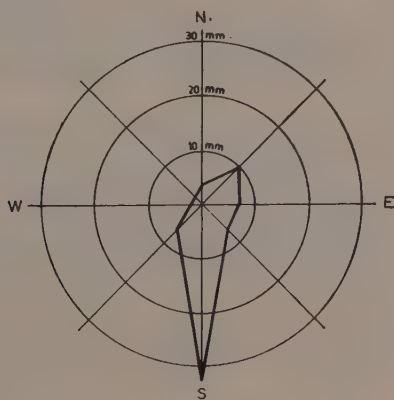


Fig. 4

Fig. 4 shows the distribution of the amount of precipitation during different prevailing wind directions at time of snow fall up to January 15, at snow station in Juuka during the accumulation of snow in winter 1959-60. It is seen that the greatest amount of precipitation fell during prevailing southern winds. This calls forth computations on the influence of one individual pine on the accumulation of snow cover in north-south direction. Employing in the computations two different methods of grouping trees, two different results are obtained, which in Fig. 5 are marked with dots and crosses respectively.

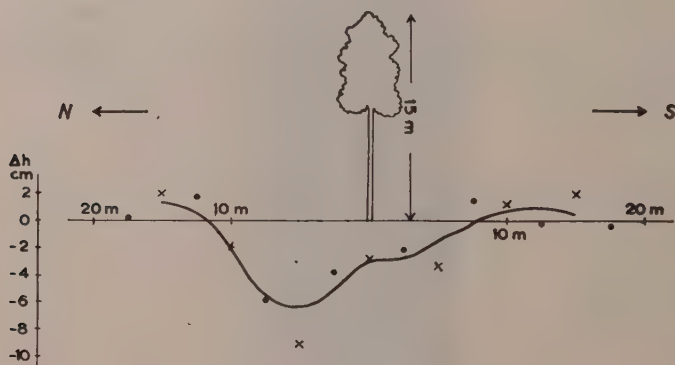


Fig. 5

Fig. 5 shows the influence of one individual tree on accumulation of snow in pine dominated forest in north-south direction at snow station in Juuka on January, 15, 1960. The vertical profile area of the pine is 25 square meters, including trunk, and the height of the tree is 15 meters. The point of least depth of snow falls averagely so close to the tree that one may take for granted the trunk of the tree to have had a considerable influence in forming this minimum.

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DEVELOPMENT OF SNOW STUDIES IN THE USSR

G. RICHTER

U.S.S.R. Academy of Science Institute of Geography

SUMMARY

1. Foundations in the science of snow have been laid down in 1871 by the Russian scientist A. I. Voeikov, but it was actually developed only within the last decades, owing to the extensive construction projects, which embraced all the regions of the Soviet Union.

2. In connection with the reconstruction of waterways and the construction of hydroelectric power plants, researches of laws governing the distribution of snow reserves, melting processes and run-off of meltwaters have been organized on a wide scale.

3. An extensive use of snow in agriculture and the fight against snow drifts on communications, demanded a study of the processes of snow transport and accumulations, as well as the establishment of its physical and mechanical properties on a regional scale. A wide application of snow retention and other methods of snow amelioration (improvement of water-heat regime of the soils by regulating the accumulation of snow) was facilitated by the state monopoly on land in the USSR, which permitted to carry out snow amelioration on vast areas and to mechanize this labour-consuming process.

5. Great attention is paid to the studies of snow because of its tremendous role in all natural processes (formation of local climate, development of relief, soil formation, distribution and biology of plants and animals). Many papers have been devoted to this problem.

RÉSUMÉ

1. Les éléments de la science de la neige furent fondés par le savant russe A. Voeikov, mais elle se développa seulement au cours des dernières dizaines d'années en rapport à la construction immense, qui embrassa toutes les régions d'U.R.S.S.

2. En relation à la reconstruction des voies d'eau et à la construction d'hydrocentrales furent organisées des investigations de grande proportion sur les lois gouvernant la distribution des réserves de neige, les processus de fonte et d'écoulement des eaux de neige.

3. La vaste utilisation de la neige en agriculture, et la lutte contre les enneigements des voies de communication ont provoqué la nécessité d'étudier les processus de translation et d'accumulation de la neige et de ses qualités physico-mécaniques sous un aspect régional. A la vaste application de retenue de neige et à d'autres méthodes de bonification de neige (amélioration de régime thermo-aqueux des sols par réglage des accumulations de neige) contribua la socialisation en U.R.S.S. des terres; ceci permit de procéder aux bonifications sur des vastes étendues et de mécaniser ce processus, qui exige beaucoup de main-d'œuvre.

4. Une grande attention fut payée aux problèmes de l'étude de la neige en rapport à son rôle immense dans tous les processus naturels (formation du climat local, développement du relief, distribution et biologie des plantes et des animaux), auxquels sont dédiés maints ouvrages.

The wide stretches of temperate and polar zones of the globe, as well as the high mountains abounding in all zones of the earth down to the Equator, are covered with snow layers of various thickness during a more or less lengthy period of time.

Owing to its high reflectivity and low heat conductivity, the snow blanket has considerable influence on the heat regimen of the atmosphere and soil as well as on their heat exchange. As a result of this, the development of natural processes in regions subjected to snow blanketing differs from that of regions deprived of snow. The snow mass protects the soil from deep freezing and from sharp fluctuations of temperature; the plants and animals hibernating under the snow enjoy more favourable conditions for enduring the dangerous frosty weather. The process of development of surface

irregularities and soil formation under snow blanketing goes on at entirely different rate of speed. Atmospheric precipitation accumulated during the winter under snow cover has great influence on the feed and regimen of surface and underground waters. In mountainous countries snow constitutes the principal source of glacier feeding.

The snow blanket represents an original rock formation, consisting of ice crystals, interspaced with air, containing vapourized or liquid water. Under the influence of temperature fluctuations, moisture and pressure, the proportion of solid, liquid and gaseous water particles in snow constantly changes, causing, in turn, the changing of numerous properties of the snow blanket and of its influence on natural processes. The extent of influence exerted by the snow blanket on nature depends upon the regimen of the snow mass and particularly upon the time length of its occurrence.

In the economic activity of man the snow blanket plays no less a part, than in nature. Snow has a particular significance in agriculture, where the harvest on vast extents of territory is in direct dependence upon the quantity and character of snow occurrence. In the steppe and tundra regions, which are open to snow storms, there is an annual heavy transfer of snow, encumbering, as a result of snow-drifts, the work of automobile and railroad transportation. This leads to large expenses and loss of working energy used up to clean away and protect the communications from snow obstructions. In mountain regions the accumulations of snow cause catastrophic snow avalanches (snowslides) of great destructive power. In regions of prolonged snow occurrence it is used as building material for the construction of snow-and-ice roads, of temporary huts and snow-ice refrigerator warehouses.

Despite its wide-spread distribution over the surface of the earth and the large part played in nature and human economic activity the snow blanketing has only recently received close examination. The pioneer in the field of snow studying was the prominent Russian climatologist and geographer, professor A. Voeikoff. He was first in the field of scientific literature to draw attention to the necessity of snow courses in a short article entitled: "The influence of snow-covering on the climate", published in 1871. Later on, in 1885-1889 he published more extensive investigations in this field, which retain their interest until the present day and have marked the beginning of a new branch of science on snow, or *snow courses*.

A. Voeikoff was the first in the world to organize a net of investigation points for the study of snow accumulations, which were subsequently transferred to the control of meteorological stations supervised by the Main Physical Observatory. The observation data collected by these meteorological stations were digested and published by A. Voeikoff in the magazine "Meteorological Herald", and in other publications. A. Voeikoff wrote as far back as 1889: "Observations of the snow blanket are so important, that it would be advisable to raise the question about them at the next International Congress of Meteorology. It would be best that it should be raised by the Russian representative, since nowhere is the territory annually covered by snow as extensive as in our Motherland." A. Zupan, the well-known German geographer, deservedly attributed to A. Voeikoff the name of the "father of snow courses".

One of the main problems of snow courses is to determine the reserves of water in snow, which is of great practical significance for hydrological forecasts. Prior to the Revolution the net of meteorological stations was sparse and observations did not ensure reliable hydrological forecasts. Furthermore the methods of observations carried out on solitary gauges did not ensure sufficient accuracy. At the present time the net of meteorological stations has increased many times and has begun to cover the entire territory of the USSR at a more uniform rate. Considerable improvement has been reached in the methods of observations and calculation of the water reserves contained in the snow blanket.

Courses on snow accumulation and snow thawing have received particular development during the years of rapidly growing national economy in the USSR and

vast hydrotechnical construction. The erection of numerous hydrotechnical stations on the Volkhov, Svir, Dniepre, Volga, Angara, Yenisey, Obj and many other rivers, the construction of large canals (the Belomor-Baltic, Moscow-Volga, Volga-Don, etc.) and of irrigation and drainage systems require the preparation of precise hydrological estimations and forecasts. Since the majority of USSR rivers are fed more than 50% by snow melt waters, the problem of studying snow accumulation, snow thawing and run-off of melt waters became especially urgent. In various scientific institutes including the State Hydrological Institute, the Central Forecasting Institute, the Main Geophysical Observatory, the USSR Academy of Science Institute of Geography and many others, wide investigation of the processes of snow accumulation and snow thawing is carried on, considering the different weather conditions, surface irregularities, and plant species; methods are being developed for hydrological forecasts and estimations of snow floods (conducted by G. Kalinin, P. Kuzmin, V. Komaroff, V. Rakhmanoff, M. Sribnyi and others), helping to ensure the necessary accuracy of technical estimations. The newly developed method of estimating snow thawing with the aid of the heat balance equation finds widespread application.

The necessity of studying the physical and mechanical properties of snow, and particularly of investigating the snow transfer by wind, was encountered during the construction of the first railways, when the problem of snowdrifts impairing normal transport conditions had to be coped with. As far back as 1863 the engineer V. Titoff first recommended the method of coping with the problem of snow-drifts by setting up portable wooden guards. In order to estimate the design and application of guards, thorough field and laboratory investigations had to be made on the snowdrift flow, the distribution of snow in it and the influence of mechanical obstacles of various construction on the flow structure and on the snow drifts (N. Zhukovsky, N. Dolgoff and others). The designs of portable wooden snow guards proved quite effective and their widespread application on railways practically eliminated the snow drifts on Russian railroads. However, railway guarding involved considerable expenses in money and human energy. To take the place of portable guards came the less costly hedge of bushes and trees planted along railway lines. Such protection effectively retained the passing snow, keeping it away from the road bed. At the present time extensive work is being done in selecting trees and bushes, growing in various climatic zones, and in designing belts and hedges, which would replace throughout the wooden guards. Research work in this direction assumed a large scale in recent years, due to the increase of new railway and automobile road construction. Alongside with this work, investigation is widely conducted in the field of studying the theory of snow transferring (A. Djunin, A. Komaroff, D. Melnik and others).

The measures adopted for protecting roads by retention of transferred snow and accumulating it near the road bed involve the danger of road damage by melt waters (washing away of the road bed, landslides, etc.). In order to avoid large accumulations of snow near the road bed, snow retention is being practised in the place of its falling, thus reducing the total mass of transferred snow. This method has acquired the name of "complex" method, since it not only protects the road bed from snowdrifts, but also tends to increase the reserves of snow on the adjoining fields, where this snow serves as an additional means for soil moistening, intended to improve agricultural crops. Particular significance is gained by this method in steppe and dry zones suffering from inadequate natural soil moisture.

In mountain regions, owing to accumulations of large masses of snow brought by the winds, huge avalanches are formed, which attract the attention of numerous investigators (V. Akkuratoff, G. Sulakvelidze, G. Tushinsky and others). In examining these avalanches not only the snowslides themselves are investigated and the dangerous avalanche zones are established, but also the structure of the snow mass, the processes of snow metamorphosis and the entire complex of causes leading to

slides are studied, to aid in the foretelling and preventing the occurrence of avalanches. In some mountain regions special stations are created for studying and forecasting avalanches and working out methods for combating them.

Although snow and melt waters have been utilized since times immemorial to raise agricultural crops, the first references in literature to the possibility of combating droughts through snow retention were made only in the beginning of the XIXth century (V. Lomikovsky 1837). In the works of many prominent Russian scientists of the end of the XIXth century (N. Adamoff, A. Voeikoff, N. Vysotsky, V. Dokuchaiev, A. Izmailsky, P. Kostychev and many others) there are numerous methods recommended for the utilization of snow to raise crop yields. However, prior to the Revolution the development of snow retention practice was hindered by private ownership of land and by the petty parcellization of fields. It was only after collective agriculture was introduced and large collective farms were organized, that snow retention work could be enforced and mechanization applied on uninterrupted stretches of fields.

By utilizing the mobility of snow and its changeable properties induced by artificial redistribution, reflectivity, density and heat conductivity, it became possible to improve the thermal and water regimen of the soil, to prolong or shorten the vegetation period, to control the intensity of snow thawing and surface run-off of melt waters. All these methods tending to improve the growth of cultivated plants have received the name of "snow amelioration".

The vast material accumulated in the field of investigating and utilizing snow has found its generalization in numerous summarized publications. After the works of A. Voeikoff in 1932, was issued the book of Professor P. Chirvinsky on "Snow and Snow Retention", where snow was treated like a mineral and the snow blanket, as a rock substance, and description was given of the snow surface forms, the methods of snow utilization in agriculture and of how to combat snow drifts. The physical and mechanical properties of the snow blanket were the subject of the special summaries published by B. Weinberg (1940), G. Richter (1945), P. Kuzmin (1957), P. Shumsky (1955). The collected works of G. Richter (1948) are devoted to the role of the snow blanket in the physico-geographical process, while those of A. Formozoff (1946) and A. Nasymovitch (1955) are devoted to the special problems of the snow blanket influence on the life of mammals. G. Tuchinsky (1949, 1957) in his summaries works treats on the problems of the formation and slide of avalanches, while the numerous summaries of F. Antonoff (1951), G. Bjalobzhesky (1955, 1956), A. Kungurtseff and P. Sarsatskikh (1960), of D. Melnik (1955), and G. Tushinsky (1960) deal with the problems of snowdrifts. If during the primary stage of snow blanket courses principal attention was paid to questions of quality, the works of the recent period show that the investigators pay most attention to the quantitative evaluation of events and processes.

Literature on snow course in the Soviet Union is increasing from year to year. During recent years the USSR Academy of Science Institute of Geography systematically publishes summary works on the subject of studying and utilizing the snow blanket. Apart from these summaries, various magazines and publications annually publish hundreds of articles on various problems connected with snow course. Despite this vast amount of literature, our knowledge of snow is still far from perfect and is incomplete. A huge amount of work is still to be done. The main difficulty in snow course lies in the exceptionally unstable character of its properties, changing under the influence of surrounding conditions.

Up to the present time snow course activity was uncoordinated and assigned to specialists of different branches: falling snow, as one of the varieties of atmospheric precipitation, was the task of meteorologists, the distribution of water and snow reserves—was the task of hydrologists, individual physical and mechanical properties

were assigned to specialists depending upon their practical branches of physics, etc. This disconnection of problems for investigation produced a lack of uniformity in the development of separate problems constituting part of an integral problem. The treatment of the entire complex of problems connected with snow as a special formation of nature, requires unification and coordination of efforts on the part of various specialists. It is for this purpose that, since many years ago, the USSR Academy of Science has been holding systematic joint coordination commissions, where the results of investigations are summarized and future plans and methods of work are discussed.

Necessity is now ripe to segregate the entire complex of problems for snow studying into a special branch of science- *snow course*.

The problems entering into this new branch of science include the all-sided studying of snow from the following points of view:

1) snow as a mineral (the crystallographic optic and physical properties of snow-flakes) i.e. *the mineralogy of snow*.

2) snow as a rock substance (its structure, physical and mechanical properties, metamorphosis, etc.) i.e. *the petrography of the snow blanket*.

3) The distribution and regional peculiarities of the snow mass, as well as the interaction of snow with natural conditions, i.e. *the geography of snow*.

The scientific data obtained is used for practical purposes in order to develop the methods of snow amelioration and most effective utilization of snow in agriculture, as well as for working out methods of rational combating snowdrifts and avalanches and to utilize snow as building material.

The new science, which has come to life in Russia—*snow course*, and has as its purpose the thorough investigation of snow and snow blanketing with the aim of a widespread and complete utilization of its properties for the development of national economy, will find the most favourable conditions and great prospects in the Union of Soviet Socialist Republics.

G. RICHTER

NEW LABORATORY AT THE SNOW EXPERIMENT STATION OF THE RAILWAY TECHNICAL INSTITUTE

M. SHODA (Japan)

SUMMARY

The present paper gives an outline of the new laboratory house of the Snow Experiment Station, Shiozawa, belonging to the Railway Technical Research Institute, the Japanese National Railways, which was put into operation from the last winter.

Since the inception of the station in 1947, some research studies on snow have already been carried out for the purpose of rationalizing the countermeasures against snow damages to railway installation, but the building was only a small, wooden one and had no cold room. The rebuilding project of the new laboratory house, equipped with two cold rooms with air circulation (-50°C : $4 \times 4 \times 2.3 \text{ m}^3$ and -10°C : $9 \times 5 \times 3 \text{ m}^3$) and a snowfall orifice ($3.5 \times 3.5 \text{ m}^2$) with an openable roof to make precipitation available for research at the snowfall room, was started in October 1957 and brought to completion in December 1958.

1. HISTORICAL SKETCH

It was in the year of 1946, immediately after World War II, that the Snow Experiment Station, Shiozawa belonging to the Railway Technical Research Institute was originally planned with the object of making research for rationalization of the countermeasures against snow damages to railway installations. At first the scope of research was confined to the railway electrification in the snowy districts of Japan.

The original scheme of the station conceived was a staff of 10 research members and 10 supporting technicians and a four-storied building equipped with a cold room and an underground corridor for observation. This however fell through for budgetary reasons and instead only a small, wooden barrack was built in December 1947 at Shiozawa, a little town situated near the centre of the main island of Japan. We were obliged to be content with such a temporary station, staffed only with a few members, inconvenient, dirty and without any cold room. Under these unfavourable conditions we have carried out various research works as follows;

Weather observations	1947 — continued
Snow accretion on wires.	1949 — 1955
Sleet jump of overhead lines.	1951 — 1955
Construction of snow cover strata.	1950 — continued
Subsidence force of snow cover	1950 — continued
Stress and load distribution of snow	1957 — continued
Snow pressure of sloping snow cover	1954 — continued
Impact force of falling snow mass	1955 — 1957
Artificial release of avalanche.	1957 — continued
Model test of sloping snow cover	1957 — continued

In the meantime the original plan had been reconsidered over and over again until the executives of the Japanese National Railways reached the final decision to enlarge the station. The rebuilding project started on October 1, 1957 and was brought to completion in December 1958.

For the foundation and development of the station, we owe utmost to the unchanging and warm support of Prof. U. Nakaya, Dr. S. Seki (Director of the Japanese

National Railways), and many others, as well as to the kind advices of Dr. H. Bader, Chief Scientist of the Snow, Ice and Permafrost Research Establishment of the United States of America, and of members of the staff of the Building Research Division of the National Research Council of Canada, which enabled us to draw up our concrete plan of the cold rooms.

2. NEW LABORATORY HOUSE



Fig. 1

2.1. Two cold rooms

	Cold room no. 1 (−50 °C) 4 × 4 × 2.3 m ³	Cold room no. 2 (−10 °C) 9 × 5 × 3 m ³
<i>Insulation</i>		
Ceiling and wall floor	20 cm, foamstyrene 25 cm, «coapolite»	10 cm, foamstyrene 15 cm, «coapolite»
<i>Vapor-seal:</i>	roofing papers and asphalt	
<i>Refrigeration equipment</i>		
1st-stage compressor	Hitachi, 3-cylinder 15 hp motor	Hitachi, 4-cylinder 15 hp motor
2nd-stage compressor	Hitachi, 4-cylinder 15 hp motor	
<i>Lowest temperature:</i>	− 60 °C	− 20 °C

2.2. Plan and section

2nd story

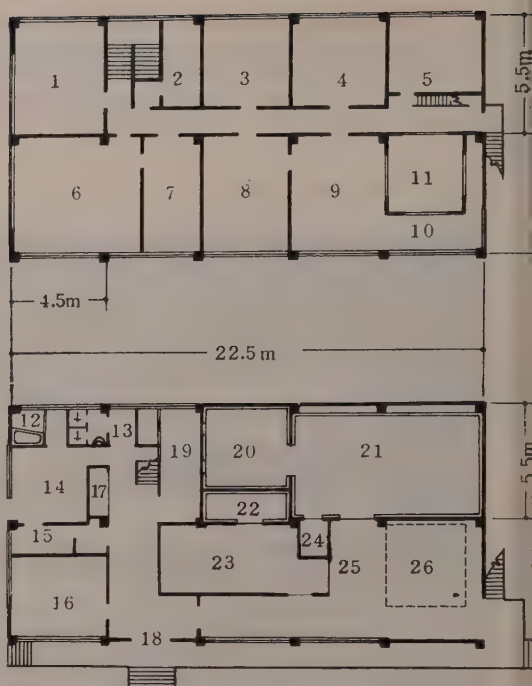
1. Resting room.
2. Darkroom.
3. 2nd laboratory.
4. Drawing room.
5. Storeroom.
6. Office.
7. Library.
8. 1st laboratory.
9. Observatory for weather.
10. Observatory for snow cover.
11. Snowfall entrance.

1st story

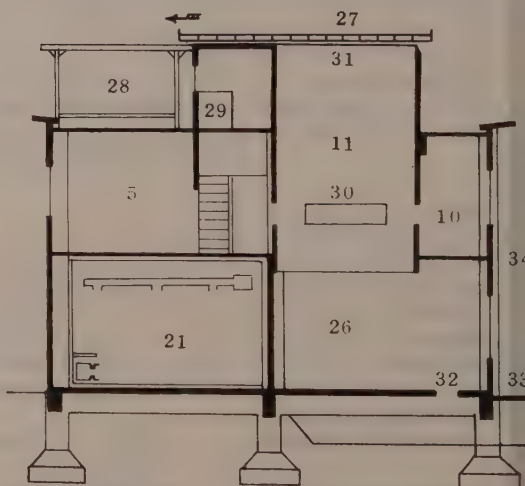
12. Bath-room.
13. Toilettroom.
14. Waiting room.
15. Kitchenette.
16. Workshop.
17. Gas storage.
18. Porch.
19. Observatory for cold room.
20. Cold room no. 1, -50°C .
21. Cold room no. 2, -10°C .
22. Diffusers.
23. Machinery room.
24. Airlock.
25. Control panel.
26. Snowfall room.

Section of the building

27. Openable roof.
28. Roof.
29. Machine to move roof.
30. Windows for observation.
31. Orifice.
32. Snow dump holes.
33. Terrace.
34. Facade.



Plan of the new station



Section

Fig. 2

2.3. Photographs reproducing

(1) Openable roof, movable by pressing a button at the snowfall room to make precipitation available for research there as much as we want through the orifice 3.5×3.5 m.

(2) The temperature indicator and controllers for the rooms and the doors as viewed from the snowfall room².

(3) Observatory for snow cover, overlooking the outdoor apparatuses such as the level bars; we can measure (subsidence force acting on the bars, snow load on the ground ...), temperature distribution in snow cover, etc. in this room.

(4) Three refrigeration compressors, employing Freon-22 refrigerant. Left is the one for -10°C room, and right two are for -50°C room.

(5) Cold room no. 1 (-50°C).

(6) Cold room no. 2 (-10°C); cooled with air circulation. Air is cooled in a unit cooler, placed in a room behind the left side wall, comes into the room through the overhead duct and goes out through the punched plates under the table (right side) to return to the cooler again. The centre is the door of another room (-50°C).



Photo (1)



Photo (2)

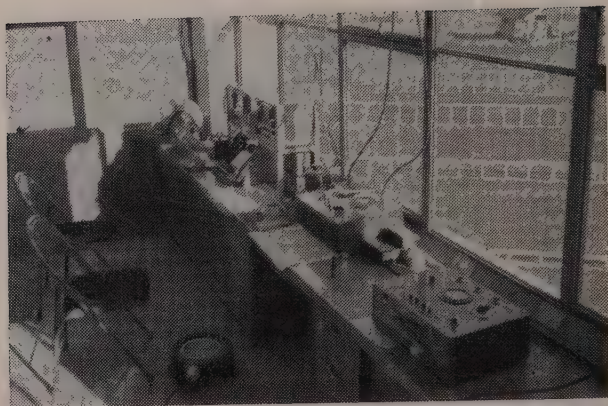


Photo (3)

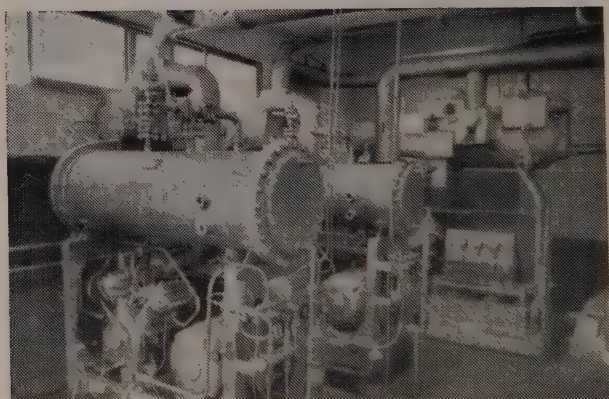


Photo (4)

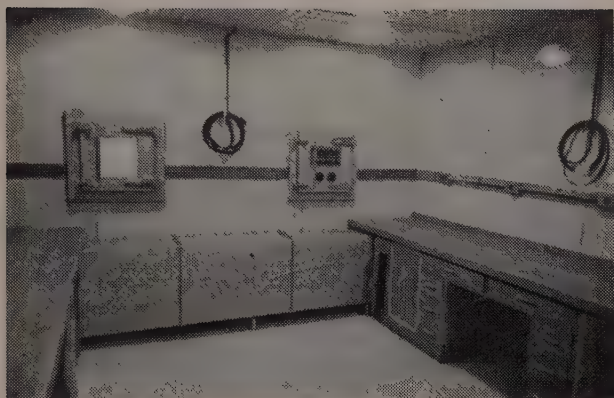


Photo (5)



Photo (6)

WIND TUNNEL STUDIES WITH SCALE MODEL SIMULATED SNOW

R.W. GERDEL* and GORDON H. STROM**

SUMMARY

In Polar regions where little or no summer melting occurs, improperly designed structures may be quickly and permanently buried by drifting snow.

In most wind tunnel studies on drifting snow no consideration was given to the relationship between the velocity of air in the tunnel and the physical and aerodynamic properties of the material selected to represent snow nor to the extent of saturation of the wind with the synthetic snow.

Recognizing the deficiencies in knowledge on snow drifting and the advantages inherent in wind tunnel studies, the U.S. Army Snow Ice and Permafrost Research Establishment has supported a research program leading to the selection and use of materials which might be used to suitably simulate snow in controlled investigations on scale models of structures within the range of 1/10 to 1/50 prototype size.

Some of the results of the wind tunnel studies with a scaled, simulated snow are presented in this paper.

1. INTRODUCTION

In the Polar regions the surface snow is in almost continuous movement. On the Polar Ice Caps, such as those covering Greenland, portions of the Canadian Archipelago and the Antarctic Continent where little or no summer melting occurs the persistently drifting snow interferes with both ground and air transportation, creates major maintenance problems where facilities must be kept operational and may quickly and permanently bury important installations. From their studies of drifting at Maudheim, the headquarters for the Norwegian-British-Swedish Antarctic Expedition of 1949-1952, Roots and Swithinbank (1954) concluded that by proper layout and construction of an Arctic base many of the operational problems caused by drifting snow could be eliminated or greatly alleviated. Finney (1939) and Rikhter (1945) have shown that control of drifting snow may be achieved by proper highway design. Pugh and Price (1953) and Baughman (1958) have reported on field studies of snow drift control by means of snow fences. Gerdel (1958) showed that where the prevailing storm winds are persistently from one direction a properly designed pattern of drift fences may be used to accumulate a large, oriented snow drift that can be processed into an elevated landing strip for air craft which might be serviceable for several years with only minimum interference by snow drifting and snow removal operations.

Rapid expansion in the construction of installations for military and scientific research in the Arctic and the need for specialized structures of untested, unique design prohibits the conduction of field tests under natural conditions with full scale models. Field tests are not only time consuming and expensive but due to the vagaries of natural phenomena it is almost impossible to duplicate a test or to evaluate modifications of a prototype structure under identical test conditions.

Satisfactory control of the environment can be attained only by the testing of suitable scale models in a wind tunnel. In a tunnel with a properly designed test

* Chief, Climatic and Environmental Research Branch, U. S. Army Snow Ice and Permafrost Research Establishment, Wilmette, Illinois.

** Director of Wind Tunnels and Professor of Aeronautics, New York University, New York, N. Y.

chamber realistic scale models can be studied, modifications made and the test repeated until the most satisfactory design is found. Finney (1939) used wind tunnel test to evaluate highway design, embankment slopes and snow fence placement for drift control. Warnick (1956) tested models of precipitation gages and shields in a wind tunnel to determine the design most effective for use in areas of heavy snow fall. These and other investigators fail to recognize the need for scaling the properties of the materials used to simulate snow to conform with the scale of the structures being tested and the physical dimensions of the test chamber and aerodynamic characteristics of the tunnel.

2. PROBLEMS ASSOCIATED WITH WIND TUNNEL STUDIES

Scale models may be so small that they project entirely into the stream of moving simulated snow in a wind tunnel whereas under natural conditions in the field most of the elevation of the prototype would be above the layer of blowing snow. This is particularly true when a low density material such as sawdust is used to simulate snow and test model dimensions are in the order of 1:25. The elastic properties of the snow simulating materials may be, as in the case of sawdust, much greater than for natural snow, consequently the relationship between the amplitude of rebound of the simulating material and the dimension of the scale model being tested may be unrealistic.

The effect of large differences between the elasticity of a simulating material such as sawdust and natural snow is shown in Figure 1. These photographs taken in

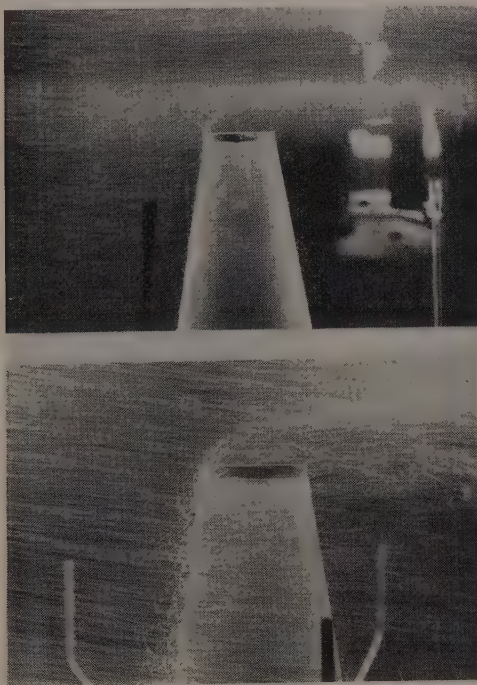


Fig. 1 — Showing the difference in elasticity between natural snow and sawdust which was used at the University of Idaho to simulate snow in wind tunnel tests on precipitation gages. The upper picture is of natural snow, the lower of sawdust. (From Warnick op cit).

the wind tunnel at the University of Idaho (Warnick (1956)) effectively illustrate how among other properties to be scaled the coefficient of restitution, which is the ratio of the velocity of rebound to the velocity of impact must be included in the selection of materials for the modeling of drifting snow.

The complex nature of snow and of the phenomenon of drifting makes the determination of modeling criteria for a simulated snow very difficult. Fortunately certain geometric properties which would at first appear to be insurmountable become of little importance when dealing with blowing snow on the Polar ice sheets. Under continuous movement by the wind the dendritic and spicule forms of snow flakes are rapidly abraded into rounded grains and, except for some small air inclusions assume most of the characteristics of ice grains. Further, although the mean size of the grains may vary from 0.4 mm to 1.0 mm or occasionally even larger, the particle size distribution of the snow grains in a specific geographic or climatological area covers a very narrow range, frequently not exceeding $\pm 15\%$ of the mean. The simplification of some of the geometric parameters alone greatly expands the fields of materials which may be adaptable to the simulation of snow. There are, however many physical properties which must be scaled in any really satisfactory model snow.

The size and shape of wind tunnel test section must be considered also. There is some question whether velocity profiles produced over an almost limitless distance on the earth's surface can be properly scaled in a wind tunnel test section. The boundary layer in a wind tunnel is wedge shaped having little or no thickness at the beginning of the test section and increasing in thickness downwind. Such a condition does not exist on a relatively smooth ice cap where there is usually a uniform boundary layer several hundred feet thick. With a sufficiently long wind tunnel test section the experiments may be performed at a downstream location where the rate of growth of the boundary layer and change in shear stress are at a minimum. However, realistically proportioned test models may project through the boundary layer in most tunnels suitable for simulated snow studies.

Recognizing the advantages inherent in wind tunnel testing of snow drifting the U.S. Army Snow Ice and Permafrost Research Establishment has supported a program of research in the wind tunnels at New York University, College of Engineering, Research Division.

3. CRITERIA FOR A SNOW SIMULATOR

In order to select a suitable snow simulating material the following scale factors were considered significant.

$$\frac{d}{L}, \frac{V_p^2}{gd}, \frac{V_p}{V}, \frac{V_f}{V}, e$$

where :

L = linear reference dimension of rigid boundary objects such as buildings and snow fences.

d = diameter of snow particle

V_p = velocity of snow particle

V_f = free fall velocity of snow particle

V = Ambient air velocity at the particle

g = acceleration due to gravity

e = coefficient of restitution

In application, the second and third factors may be combined to form the Froude number based on fluid velocity V^2/gd . The density of the model particule must be such as to satisfy both d/L and V_f/V .

The wind tunnel at the New York University has a test section 3.5 feet high and 7 feet wide with a usable length of 30 feet. Provisions were made to introduce the simulated snow through a slot in the ceiling at the forward part of the test section. The first models to be tested were to be constructed on a 1:10 scale, a convenient size for the test section and providing a realistic scale which would permit incorporation of irregular features in the test model. The determination of the essential properties for the simulated snow to be used in the first series of wind tunnel studies was based upon the 1:10 scale models that would be used.

From a series of test the free fall velocity of snow particles of 1.0 mm diameter was found to be 200 cm/sec. The coefficient of restitution e , of ice particles was found to be 0.555. This may be somewhat high since snow particles have more entrapped air than ice particles prepared from artificial ice.

The following relationship expresses the requirement for geometric similarity of particle size in accordance with the above scale factor.

$$\begin{aligned}d_m &= d_a (L_m/L) \\ &= (1/10) d_a \\ &= 0.1 \text{ mm}\end{aligned}\tag{1}$$

where the subscripts "m" and "a" indicate model and atmospheric counterpart, respectively.

Using the Froude number scale factor to form the relationship between model airstream velocity and prototype wind velocity the following equation is formed, (assuming constant g).

$$\begin{aligned}V_m &= V_a (L_m/L_a)^{1/2} \\ &= V_a (1/10)^{1/2} \\ &= 0.316 \sqrt{V_a}\end{aligned}\tag{2}$$

In accordance with the scale factor on fall velocity the same relationship applies.

$$\begin{aligned}V_{fm} &= 0.316 V_{fa} \\ &= 0.316 \times 200 \\ &= 63 \text{ cm/sec}\end{aligned}\tag{3}$$

The scale factor on coefficient of restitution requires the following equation.

$$e_m = e_a$$

Preliminary calculations showed that the simulating snow material must have a density of 2 gm/cm³ or more to give the required fall velocity of 63 cm/sec and diameter of 0.1 mm. Various materials were studied for their suitability and borax (Na₂B₄O₇·10H₂O) was found to be the most promising. Fall velocity test showed $V_f = 60$ cm/sec for borax particles of 0.1 mm diameter. Tests on coefficient of restitution showed a value of 0.334 which is lower than that determined for ice particles. This lower value may result in some suppression of the saltation phase of the simulated snow movement.

Commercial borax available in quantities sufficient for these experiments had a mean particle diameter close to 0.2 mm. This influenced the free fall velocity as well as the linear scale of particle size. The first series of experiments were conducted, however, with the available larger particles as a means of solving some of the tunnel operation problems.

No attempt was made to scale the velocity profile in these experiments other than that which resulted from the flow around modeled objects. The modeled objects were placed approximately 15 feet downstream of the location where simulating snow was dropped into the airstream. Before an experiment was started, the floor of the test section was covered with a layer of simulating snow which extended about 13 feet



Fig. 2 — A photograph of a portion of the test chamber in the wind tunnel at New York University. The useable section is 7 feet wide, 3.5 feet high and 30 feet long. The model building being studied is approximately centered in the test section.

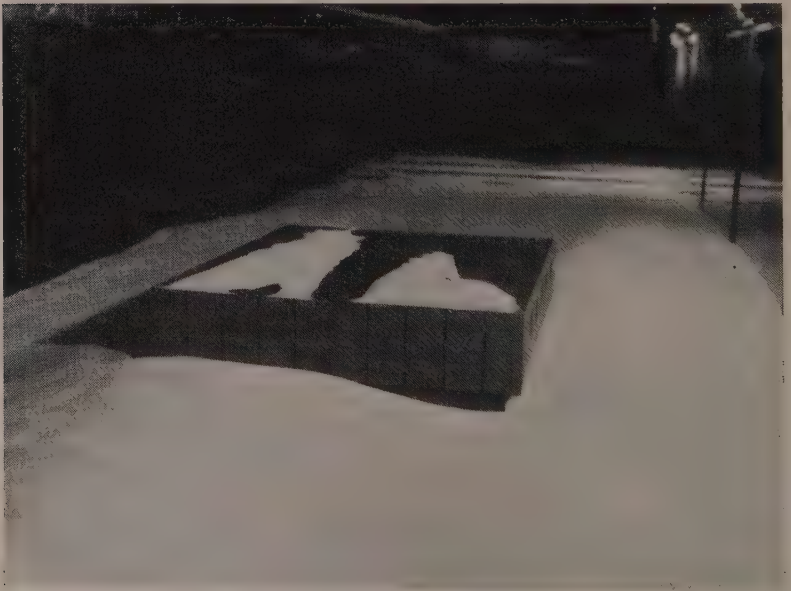


Fig. 3 — A photograph through the long axis of the test chamber showing the placement of a building model under test.

upstream of the modeled objects. Upstream of the simulated snow there was 7 feet of bare test section floor surface which joined the wind tunnel contraction cone. Figure 2 is a photograph of part of the test chamber of the wind tunnel during one of the test. The scale model building is positioned near the center of the test section. Figure 3 is a view inside the tunnel from the upstream end of the test chamber.

No velocity profile measurements were made during these experiments. However, from observations the boundary layer depth was estimated to be on the order of 1/2 foot at 15 feet where the models were centered. The model experiments may, therefore, be expected to give approximate modeling of velocity profile for a depth corresponding to 5 ft in the atmosphere. Above this level the model air flow is at constant velocity while in the atmosphere the velocity continues to increase with elevations up to several hundred feet.

The modeling of threshold velocity for the borax particles has not been solved. Computations indicate that the free stream threshold velocity for the 0.2 mm diameter borax particles is 19 mph, an almost impossible equivalent prototype speed. The wind tunnel experiments showed that the simulated snow began to move at free stream speeds on the order of 11 mph in the flat region upstream of modeled objects when no snow was falling and at a lower speed with falling snow. Irregularities in the surface of the simulated snow probably played an important part by raising the effective grain size. There is obviously need for more information on threshold characteristics.

4. MODEL TEST WITH THE BORAX SNOW SIMULATOR

An experiment was designed to test the effect of elevation of a large structure above the snow surface. The model used was selected from designs proposed for fully, self-contained structures to house the equipment and staff to operate certain Arctic facilities.

In Figure 4 the top photograph shows the model structure set on 10 foot high columns at the start of a test. The middle photograph shows the type of snow accumulation pattern developed during a scaled snow storm approximately equal to all the blowing snow storms occurring during one-half year at a selected location on the Greenland Ice Cap. During this test the moving air stream was kept near the saturation level for the simulated snow.

The lower photograph shows the pattern which developed when movement of the simulated snow was limited to creep and saltation to reproduce the conditions of a high wind without precipitation, blowing over a wind slabbed surface. The accumulation pattern shown in this photograph indicates that unless the columns are set very deep or are founded on well consolidated snow uncontrolled erosion beneath an elevated building might result in loss of support by some of the columns. In subsequent test methods for control of the erosion were studied and satisfactory designs developed.

Figure 5 and 6 are photographs showing tests of spacing and orientation of scale models of buildings of the Clements Panel type. The buildings in Figure 5 are oriented with their long axis normal to the wind stream and those in Figure 6 are positioned with their long axis parallel to the wind.

Figure 7 shows the results of a test on spacing of large fuel storage tanks. In some of these test the vortex flow around the tanks produced sufficient erosion to cause uneven settling of the tank indicating that piles or erosion control structures may be required where large storage tanks are erected on a permanent snow field.

A more comprehensive report on the first series of wind tunnel tests using a partially scale modeled snow is being prepared. The tunnel is being equipped to more efficiently handle the introduction, removal and recirculation of the many tons of

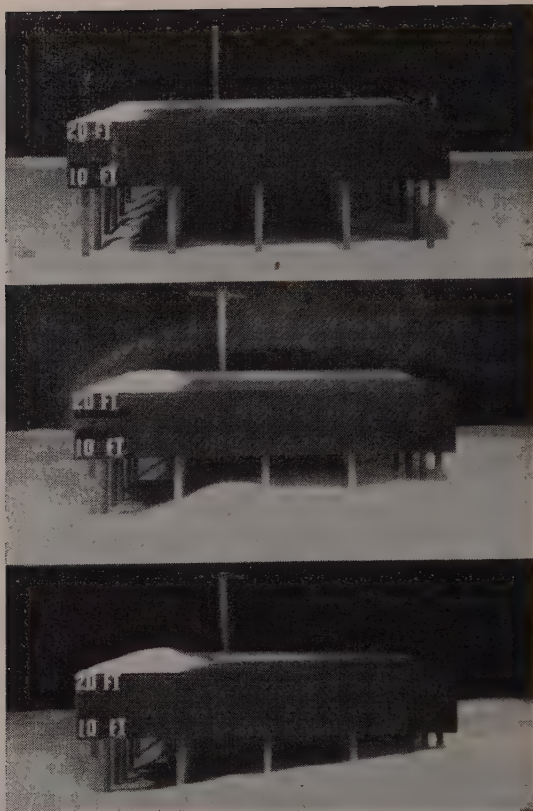


Fig. 4 — This series of three photographs illustrates a part of the test on an elevated column-supported structure. The top picture shows the model building at the start of the test. The middle picture shows the accumulation profile near the end of a heavy blowing snow storm. The bottom picture shows the accumulation profile which formed following a continuous heavy wind storm unaccompanied by falling snow when erosion due to creep and saltation removed some of the previously accumulated snow.



Fig. 5 — This photograph illustrates a test on scale models of Clements Panel buildings commonly used in the Arctic. The buildings are oriented with the long axis normal to the wind.

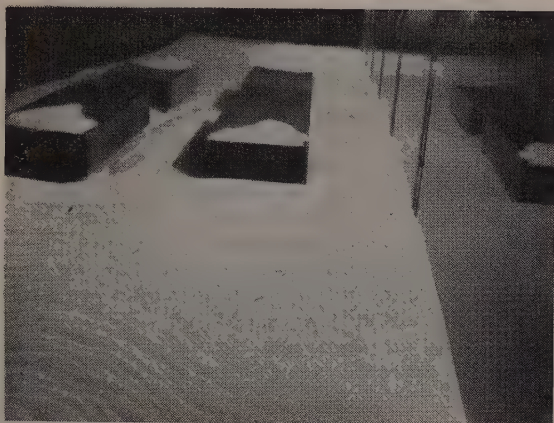


Fig. 6 — The Clements Panel buildings in this test are oriented with the long axis parallel to the wind stream.

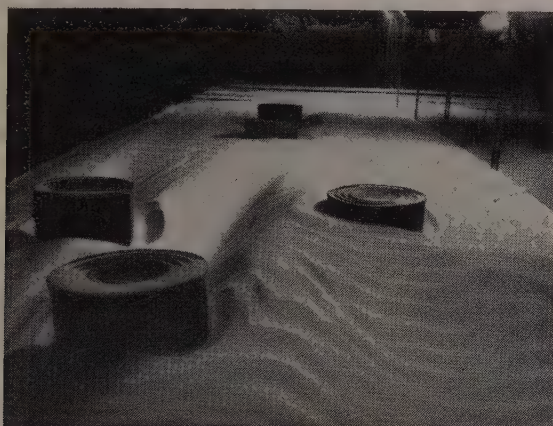


Fig. 7 — This photograph shows the accumulation pattern developed during a test of spacing of large capacity fuel tanks.

now simulating material required in the large test section. A study is under way to determine whether suitable snow simulator materials are available or can be manufactured to permit scale model testing at 1:25 to 1:50 scale so the design for a multiple building facility can be studied in the present test chamber.

DISCUSSION

Field test on the effect of blowing and drifting snow on structure design and installation layout are limited by the vagaries of weather and the many years of time required to obtain usable information.

Wind tunnel test provide a means for artificial control of the environment and

the acquirement of results in a short period of time which may be used for the development of design criteria for construction of Arctic facilities. Under suitable conditions a few hours test in a wind tunnel may provide acceptable information that could not be acquired in less than 3 to 5 years under natural field conditions.

The production of a satisfactory blowing snow environment in a wind tunnel requires that the physical properties of any material selected to simulate natural snow must be scaled in relation to the linear scale of the structure models in a manner that will insure a representative scale model environment.

A research program at New York University sponsored by the U.S. Army Snow Ice and Permafrost Research Establishment has been directed toward the computation of essential scale parameters for a model snow and the search for and selection of a material which will properly simulate snow when used with 1:10 scale models of structures and facility layouts.

The first series of test were made with borax ($\text{Na}_2\text{B}_4\text{O}_7 \cdot 10\text{H}_2\text{O}$) which of the several materials studied appears to most closely approach the theoretical requirements for a modeled snow.

Photographs showing the wind tunnel at New York University and of some of the studies conducted on structure design and layout are presented to show the test conditions. The photographs effectively show the influence of several design and layout concepts on snow accumulation.

Further studies are planned. The wind tunnel is being equipped with a conveyor system which will simplify the introduction and removal of the large volume of synthetic snow required for each test.

A search is being made for materials which will meet the requirements for a modeled snow applicable to testing at 1:25 to 1:50 scale to permit the study of layout and spacing of structures for a large and complete installation in a region dominated by blowing snow.

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THE EFFECT OF THE CHARACTERISTICS OF SNOW FENCES ON THE QUANTITY AND SHAPE OF THE DEPOSITED SNOW

W.I.J. PRICE

SUMMARY

In order to devise snow fences to protect roads from drifting snow, the British Road Research Laboratory has during recent years, carried out full-scale experiments in Scotland to see how the size and shape of the drift deposited by a snow fence is affected by the characteristics of the fence. This paper gives the results of these investigations.

The profiles of the drifts produced by the fences are described at various stages in their growth. At all stages, the density ratio (frontal area of solid part of fence divided by total frontal area) was the most important factor in determining the volume of snow deposited by a fence, the greatest volume occurring at a density ratio of 0.4. The effect of fence height (h metres) was studied over the range 1.25 to 1.87 m at a density ratio of 0.5. When the fence was saturated with snow, the maximum height of the drift was $1.14 h$ and its length was $22+6.5 h$. With wind normal to the fence, there was no significant difference in the drifts whether the slats were vertical or horizontal. The effect of slat width was studied at a density ratio of 0.42 and found unimportant over the range from 2.5 to 22.5 cm. At greater widths, less snow was deposited directly behind each slat.

When the fences were saturated with snow, the shape of the profile of the drift perpendicular to the fence could be represented graphically by part of the perimeter of one petal of a mathematical rose. The constants of the polar equation of this curve were related to the density ratio of the fence, its height, and the size and arrangement of the material in it.

RÉSUMÉ

Afin de mieux construire les pare-neige pour protéger les routes contre les congères, le Laboratoire de Recherches routières britannique a conduit des expériences en vraie grandeur pour déterminer les caractéristiques des congères formées près des divers types de pare-neige. Ce papier décrit les formes des congères formées par les divers types de pare-neige et démontre l'influence des dimensions du pare-neige sur la forme de la congère.

On décrit les profils des congères aux divers stades de leur développement; quand le pare-neige ne peut plus retenir de neige, le profil de la congère peut se représenter par une partie du périmètre d'une pétale d'une rose ayant « n » pétales. Il y a une relation entre les constantes de l'équation polaire de cette rose et le rapport de densité du pare-neige (aire de devant de la partie pleine du pare-neige divisée par l'aire de devant totale), la hauteur du pare-neige, les dimensions et la disposition des matériaux du pare-neige. Les expériences démontrent que le plus grand volume de neige se trouve près des pare-neige qui ont un rapport aire/densité d'environ 0.4.

1. INTRODUCTION

Snow fences are used to prevent the serious dislocation of transport on roads and railways caused by drifting snow (¹). The function of the fence is to produce wind eddies which cause the windborne snow to be deposited in drifts around the fence, thereby depriving the wind of its snow load before it reaches the carriageway. Consequently, the relationship between the dimensions of the fence and those of the snow drifts it creates are important in determining the most suitable construction and position for the fence.

This paper deals with the shapes of the drifts formed by the various types of fence and shows how drift shape is affected by the dimensions of the fence.

The features, which appear to have the main influence on the performance of a fence, are the following:

- 1) The inclination of the fence to the vertical.
- 2) The gap which separates the bottom of the fence from the ground or vegetation.
- 3) The density ratio, (\varnothing). This is defined as the ratio of the frontal area of the solid material of the fence to the total frontal area, including the apertures; e.g., a solid fence would have a density ratio of 1.00.
- 4) The height of the fence (h).
- 5) The size and arrangement of the material in the fence.

Field trials to investigate the influence of the features described in 3), 4) and 5) were carried out by the Road Research Laboratory at a number of hilly sites in Britain during the winters of 1951-5. The main purpose of these trials was the measurement of the profiles of the drifts as soon as possible after they were formed. This was facilitated by the use of depth gauges placed around the fences (Fig. 1). Relevant meteorological data were recorded as frequently as was possible on the sites, which could not be manned continuously. The results of these trials are discussed in this paper in relation to the findings of other investigators. This is preceded by a brief review of existing knowledge on the influence of features 1) and 2) on the performance of the fence and its relevance to the field trials described subsequently.



Fig. 1 — Site of experimental snow fences, Dalwhinnie, Inverness-shire.

1.1. *Effect of inclination of fence to vertical*

The investigations of Finney ⁽²⁾ and Hallberg ⁽³⁾ have shown that a fence may be inclined with the wind, at an angle to the vertical up to 30°, without disadvantage, but larger angles of inclination reduce its snowcollecting capacity. It is accepted in practice that, in terms of both simplicity of construction and maximum efficiency for a minimum weight of fence material, the vertical or upright fence is preferred.

Only when a moveable fence is required, in districts of relatively low wind velocity, is the inclined or trestle fence likely to show a practical advantage. In consequence, the inclination of the fence was not investigated in the trials referred to in this paper and all tests were done with vertical fences.

1.2. *Effect produced by gap between bottom of fence and ground*

A gap of 15 to 30 cm below the fence and above the ground or vegetation is desirable. If the gap is less than about 7 cm, snow accumulates against the fence and reduces its effectiveness by slowly transforming it from a bluff obstacle to a streamlined one. The pressure of the snow against the fence may also cause structural damage. If, on the other hand, the gap exceeds 30 cm, Finney⁽²⁾ has shown by model experiments, that the size of the wind shadow to leeward of the fence is reduced and this reduces the snow-collecting capacity of the fence. This effect cannot be readily checked by field trials in Britain because of the lack of precision with which the gap below the fence can be set in the hilly terrain in which snow fences are usually sited. Consequently, all the trials were carried out with a gap of 20 ± 5 cm.

2. RESULTS OF FIELD TRIALS

2.1. *Effect of density ratio (σ)*

Four stages in the growth of snowdrifts around five fences of various density ratios are drawn in Fig. 2. In all cases the fences were 1.87 cm high and were composed of vertical wooden slats. During the period when these observations were made, i.e. 26.1.54 to 12.2.54, there was a continuous accumulation of snow in the drifts. The wind was fairly constant in direction, veering between 60° and 90° to the fence line, but its velocity at 2 m above the snow surface fluctuated between 3.5 and 11.0 m/s. Maximum and minimum values of air temperature were $+3^\circ\text{C}$ and -15°C respectively.

The profiles shown for each fence are the average of those measured on three vertical sections normal to the fence line, i.e. approximately in the direction of the wind. Considerable variation in the profile occurs from one section to another at the same fence. The leeward end of the drift alternates between spur and cornice formation over short intervals, as may be seen in Fig. 1. This lack of uniformity in the profile of the drift is probably a reflection of the transverse variation of the quantity of drifting snow. This, in turn, is caused by the unevenness of the terrain upwind of the fence and the effect is modified considerably by small changes in wind direction. Thus, a two-dimensional representation of the form of the drift is not entirely satisfactory, but it is simple and sufficiently accurate to show the general variation of the profile with density ratio.

During the early and intermediate stages of growth, the profile of the drift to leeward of the solid fence, density ratio 1.00, shows two distinct discontinuities on each side of the highest point. It differs in this respect from the drifts to leeward of the other fences, with density ratios of 0.74 and less, which show only one discontinuity. Frequently, cornices develop at these discontinuities. The two discontinuities to leeward of the solid fence are associated with the two large scale eddies or vortices in the wake of this fence. The movement of the air in these eddies is rapid and well-defined. Next to the fence, the eddy is almost circular and the air moves in a clockwise direction for the arrangement shown in Fig. 2. This circulation is boosted by the flow of air through the gap below the fence. It is evident from Fig. 2, that there is little deposition of snow in this region; with an increase in the free wind velocity there may even be erosion of an existing drift. Further downwind the eddy is longer, less

intense and of elliptical section, with an anticlockwise circulation. Where the two eddies meet there is a region of comparative calm, which is approximately defined by the drift profile at Stage 1 in Fig. 2.

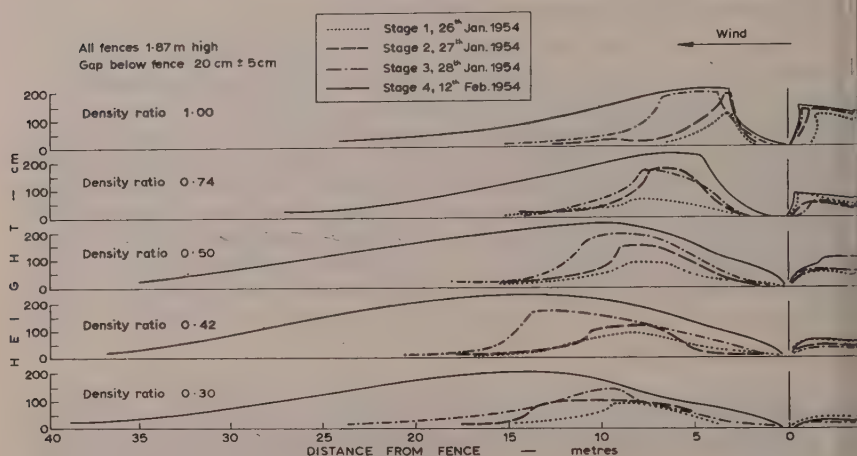


Fig. 2 — Effect of density ratio of fence on profiles of snow drifts at four stages in their growth.

This pattern of air flow within the wake of the solid fence has a powerful influence on the drift profile as it approaches its saturated form, as in Stage 4. As a result, both the windward and leeward slopes are concave upwards and there is a discontinuity in the profile to windward of the highest point. The near-saturated drift profile for a density ratio of 0.74 exhibits similar, but less pronounced, features, whereas the corresponding profiles for density ratios of 0.50, 0.42 and 0.30 are generally convex upwards and of more streamlined form. This reflects the reduction in the intensity of the eddies as the density ratio decreases. At a density ratio of 0.50, no distinct pattern of eddies can be discerned in the wake of the fence.

During the early and intermediate stages of growth, there is deposition on the windward face of the drift to leeward of all the fences except the solid one. Differential deposition on this face, due to variation in the drag velocity along it, under the combined influence of fence and drift, has led to the development of the discontinuity to leeward of the highest point. This is analogous to the formation of the slip-face on sand dunes described by Bagnold⁽⁴⁾. A steep face, often with cornice, forms to leeward of the discontinuity, creating a subsidiary wind shadow, and it is in this region that the most intense rate of deposition occurs. The steep face moves downwind during the growth of the drift and in the final saturated form, when no further deposition occurs, this face and the discontinuity disappear.

Pugh⁽⁵⁾ has suggested that the fundamental shape of the cross-section of a saturated snow drift is an ichthyoid curve, whereas Chorley⁽⁶⁾ has shown that the lemniscate loop bears a close generic relationship to streamlined forms, such as snowdrifts. It seems unlikely that a family of simple curves will provide a complete quantitative description of the forms shown in Fig. 2, in view of the discontinuity in the profile of the drifts behind the fences of high density ratios. For the more streamlined forms associated with the lower density ratios, the profile of the drift can be

represented by part of the perimeter of one of the petals of a rose having n petals, the equation of the rose in polar coordinates being:

$$r = (L - l) \cos n\theta \tag{1}$$

where L and l are the distances from the fence to the leeward and windward ends o the drift respectively. n is a dimensionless quantity defined by

$$n = \frac{\pi \cdot (L - l)^2}{8A} \tag{2}$$

where A is the cross-sectional area of the drift. The values of n for the profiles shown in Fig. 2, are 10.9, 11.8 10.9 and 17.5 corresponding to density ratios of 0.74, 0.50, 0.42 and 0.30 respectively. Allowing for the errors in the measurement of drift length and sectional area, n can be regarded as approximately constant for density ratios in the range 0.74 to 0.42, but it increases sharply for density ratios of 0.30 and less.

The length and sectional area of the snow drifts are given as empirical functions of the density ratio in Figs. 3 and 4.

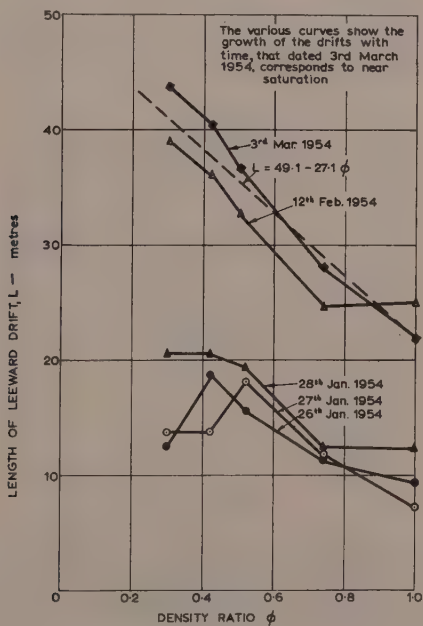


Fig. 3 — Length of drift v density ratio of fence

The drift length, L , is defined as the distance from the fence to the end of the leeward drift, measured normal to the fence line and at the level of the snow cover outside the influence of the fence. The sectional area of the snowdrift, A , is measured above the same level. This level was chosen as the lowest one at which a reasonably accurate and significant measurement of drift length could be made. Taking the observations for 12.2.54 and 3.3.54, when the sectional areas were a maximum, one of two simple expressions may be used to relate drift length, L , in metres, to density ratio, ϕ . These expressions are:

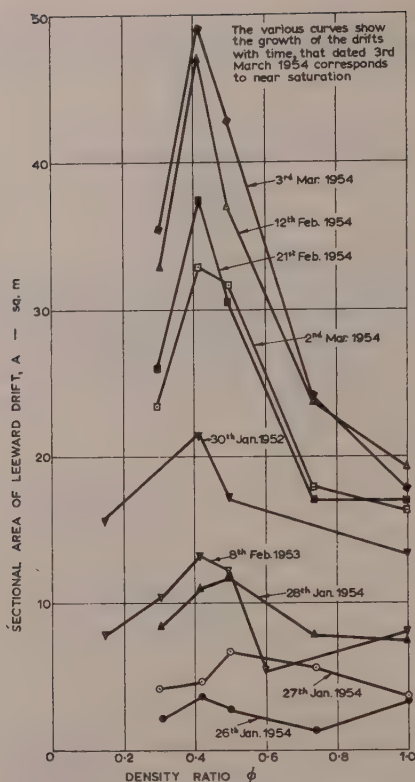


Fig. 4. SECTIONAL AREA OF LEEWARD DRIFT v DENSITY RATIO OF FENCE

$$L = 16.4 + \frac{8.4}{\phi} \quad (3a)$$

$$L = 49.1 - 27.1\phi \quad (3b)$$

Within the ranges $\phi = 0.30$ to 1.00 , equation (3b) gives the best fit, with a standard deviation of 1.6 m, compared with a standard deviation of 2.6 m for equation (3a). At values of ϕ lower than 0.2 the validity of both equations is doubtful. German investigators derived the formula (?):

$$L = \frac{15.4}{\phi + 0.22} \quad (4)$$

This corresponds to (3a) in form, but gives appreciably lower values for the length of the drift, which suggests that it refers to unsaturated conditions.

Fig. 4 shows that the sectional area of the snow drift is a maximum for a density ratio of about 0.42 . The differences between density ratios becomes more evident, the larger the drifts. The amount of snow which accumulated on the road to leeward of the solid fence was only slightly greater than that to leeward of the fence with the optimum density ratio and it was much smaller than would have been expected from

the differences in the sizes of the drifts around the fences. Similarly, the total quantity of snow collected by two fences, 1.56 m high and 30 m apart, was invariably less than the quantity collected by a single fence 1.87 m high, having the same density ratio. This implies that the protection afforded by a fence is not entirely dependent on its snow collecting capacity.

As may be seen from Fig. 2, the snow drift on the windward side of the fence does not change by very much during a prolonged period of drifting. The ratio of its area to that of the leeward drift is given in Fig. 5. When the leeward drift is approaching saturation, its height above the ground is 1.14 times the height of the fence above the ground, for density ratios of 0.42, 0.50 and 0.74. For density ratios of 0.30 and 1.00, it is 0.93 and 1.06 the height of the fence, respectively.

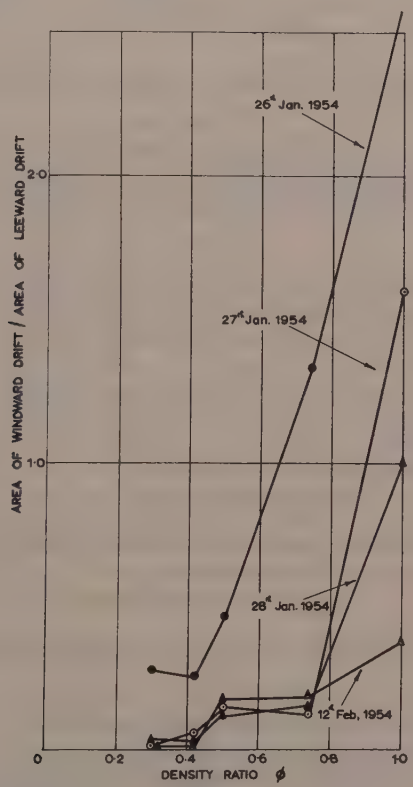


Fig. 5 — Ratio of area of windward drift to area of leeward drift v density ratio of fence.

2.2. Effect of fence height

Four stages in the growth of snowdrifts around three fences, 1.25 m, 1.56 m and 1.87 m high, are shown in Fig. 6. All these fences had a density ratio of 0.50 and a gap below them of 20 cm. They were made of vertical timber slats, of 7.5 × 2.5 cm section.

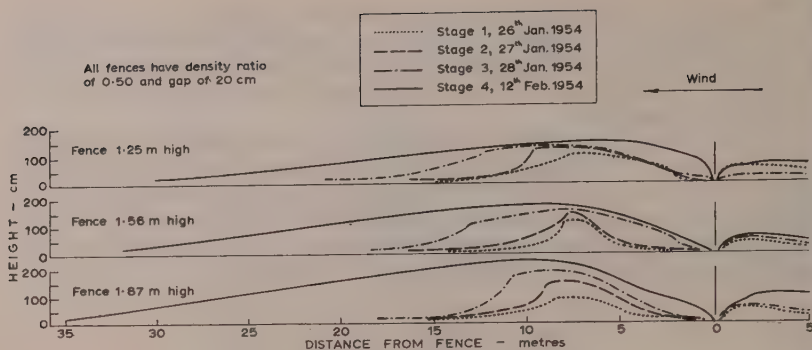


Fig. 6 — Effect of fence height on profiles of snow drifts at four stages of their growth.

It has been suggested by Nokkentved, ⁽⁸⁾ that, at saturation, the sectional area of the snowdrift, A , measured normal to the fence line, will be approximately proportional to the square of the height of the fence, h , i.e.

$$A = kh^2 \quad (5)$$

This was not found to be the case with the largest profiles measured. Average values of k were 16.4, 12.8 and 11.3 for values of h of 1.25 m, 1.56 m and 1.87 m respectively. Thus, there was not the dimensional similarity which equation (5) implies. This was reflected in the values of n , equation (2), of 15.6, 14.5 and 11.8, which correspond to the above values of h . Again, although the maximum heights of the drift were 1.14 times the fence heights, the lengths of the drifts, L were not directly proportional to the heights of the fences, h . An empirical relationship derived from the observations on 12.2.54 and 3.3.54 was of the form:

$$L = 22.1 + 6.5h \quad (6)$$

where L is in metres. This expression is similar to that given by German investigations (7), namely:

$$L = 11 + 5h \quad (7)$$

Again the values from (7) are lower than those from (6). Equation (6) may be combined with (3b) to give:

$$L = (1.38 - 0.77\phi) (22 + 6.5h) \quad (8)$$

Equation (6) refers to fence heights of 1.25, 1.56 and 1.87 m, all tested at an area density of 0.50. Equation (3b) refers to fences 1.87 m high tested at area densities ranging from 0.2 to 1. Equation (8), therefore, is not of general application but fits the lines shown in Figs. 3 and 7 as best fitting these two sets of experiments. Substituting from equation (8) for L in equation (1) and putting $l = 1$ m, gives a general equation in polar coordinates for the profile of a snowdrift in terms of the fence creating it:

$$r = [(1.38 - 0.77\phi) (22 + 6.5h) - 1] \cos \frac{20.5}{h} \theta \quad (9)$$

An empirical value of $n = 20.5/h$ is taken to define the saturation profile within the range of fence heights of 1.25 to 1.87 m. This equation is only valid for values of ϕ in the range 0.40 to 0.70, which is the range used in practice.

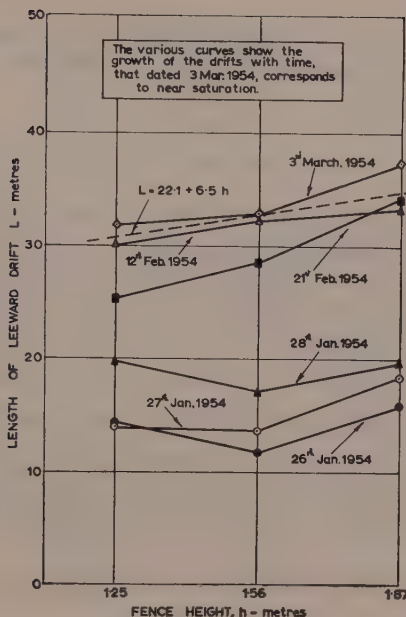


Fig. 7 — Drift length v fence height.

The absence of geometrical similarity in the profiles of drifts formed by the fences of different heights is due to two main possibilities. Firstly, the drifts around the higher fences may not have been fully saturated. Secondly, the major condition for complete dynamic similarity of the air flow in the wake of the fences is not fulfilled, in that the term e/h , where e is the roughness height of the snow surface, is not a constant, since e is the same for all the fences and h is not.

3. Effect of the size and arrangement of the material in the fence

When the snow drifts to leeward of the fences approached saturation there was no significant difference in their profiles, whether the slats were arranged horizontally or vertically, provided the wind was almost normal to the fence line. However, when the wind was inclined to the fence line at an angle less than 60° , it was noticeable that the drifts around the vertically slatted fence became shorter and steeper. This arises from the apparent increase in the density ratio due to the thickness of the slats. The thicker the slats, the greater the effect.

With fences of 0.42 density ratio, a change in the width of slats from 2.5 cm to 2.5 cm, produced no change in the drift profile. Slats 30 cm wide did show a slight shadow effect on the windward end of the leeward drift, and slats, 60 cm wide, produced a definite transverse ridge and valley profile on the windward face of the drift, up to 10 metres away from the fence. This contributed to a reduction in the total volume of snow deposited by the fence.

ACKNOWLEDGEMENTS

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NEIGE ACCUMULATION ET ABLATION
SNOW ACCUMULATION AND ABLATION

THE MECHANISM OF SLIDING ON SNOW *

MASAKI SHIMBO**

Concerning the measurement of friction of snow, model ski or sledge method has been used by many researchers (1,2,3,4,5,6) and also in our early work (7). But since these methods could not be free from front drag or scale effect, since accurate measurement of μ_S and μ_K proved very difficult (10).

To avoid this difficulty, we tried a number of methods and at last completed a new instrument (10) which has a rotational disk and a snow pan, the latter being brought into rotation by frictional sticking force between the pan and the disk.

The snow pan was tugged by spring, so that when it came to rest due to the balancing of the frictional force against the elastic force of spring, we could read the value of μ_S and μ_K directly (10).

Although the diameter and width of rotating disk involved a little scale effect, the accuracy of μ_S and μ_K measured under proper conditions could be maintained within ± 0.005 , the influence of front drag being completely eliminated (10).

Using the rotational friction instrument, we found that in the case of polytetrafluoro ethylene μ_K took constant value in the range of sliding speed 0.05 m/sec to 7.2 m/sec (Fig. 1, 2), while it increased gradually when sliding speed dropped below 0.05 m/sec (Fig. 1) (10).

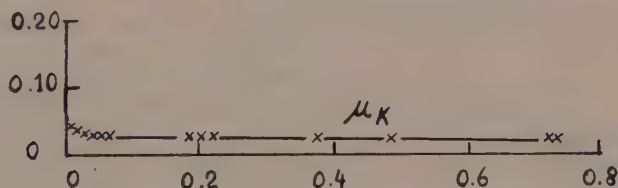


Fig. 1 — Influence of sliding speed (1) Polytetrafluoroethylene.

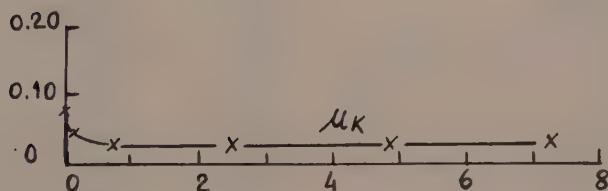


Fig. 2 — Influence of sliding speed (11) Polytetrafluoroethylene.

As to the influence of pressure on the friction of snow, we found that in the range of 21 g/cm² to 66 g/cm² μ_S and μ_K remained constant (Fig. 3) (10).

In the case of plastics μ_S was affected mainly by roughness of surfaces, and also by air temperature, snow temperature, nature and size of snow crystal, water absorption

* Abstract from paper of international association of hydrology commission of snow and ice, Helsinki meeting 1960.

** Division of research, Mizuno Co.

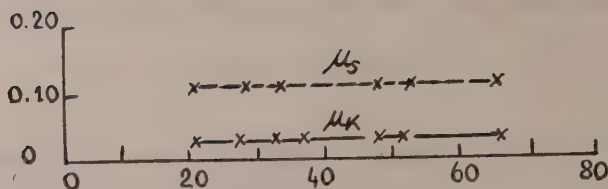


Fig. 3 — Influence of pressure—Polytetrafluoroethylene.

of plastics and water content of snow. μ_K was found to depend very little on surface roughness, but slightly on air temperature, snow temperature, nature and size of snow crystal, absorption of water by the plastics and water content of snow ⁽¹¹⁾.

The effect of air temperature on the values of μ_K is exemplified in Table 1. From such data we concluded that μ_S and μ_K took fairly large values at low temperature and when air temperatures rose close to 0°C, these values became minimum, and above 0°C μ_K of hydrophylic plastics increased suddenly while μ_S decreased, so that μ_K turned out greater than μ_S ⁽¹¹⁾.

TABLE 1

Friction of Plastics

Plastics	μ_K (-20°C)	μ_K (-11°C)	μ_K (+2°C)
Polytetrafluoro-ethylene	0.04	0.03	0.02
Polyethylene	0.05	0.04	0.03
Polypropylene	0.05	0.04	0.03
Polycarbonate	0.04	0.03	0.03
Polyurethane	0.04	0.03	0.06*
Polyamid (Nylon)	0.04	0.04	0.07*
Polyvinyl chloride	0.06	0.03	0.08*
Epoxy resin	0.05	0.03	0.07*
Celluloid	0.05	0.05	0.08*
Urea resin	0.06	0.04	0.11*
Amyno-alkyd resin	0.06	0.05	0.10*
Melamine resin	0.06	0.05	0.05*
Phenolic resin	0.04	0.03	0.07*
Styrol butadiene copolymer	0.04	0.03	0.08*
Styrol acrylonitrile butadiene copolymer	0.04	0.03	0.08*

* Abnormal increase of μ_K was observed.

Speed: 2.4 m/sec., pressure 21 g/cm².

These abnormal phenomena were observed with flat surface and excess water of wet snow but not with rough sliding surface. For example, in the case of phenolic resin and snow with low water content, μ_S increased with roughness of sliding surface

while μ_K remained constant, the relation $\mu_S > \mu_K$ holding throughout (Fig. 4). But, when cooled water was added in excess to snow, μ_K increased as the roughness diminished while μ_S decreased, with the result that $\mu_K > \mu_S$; in the region of higher

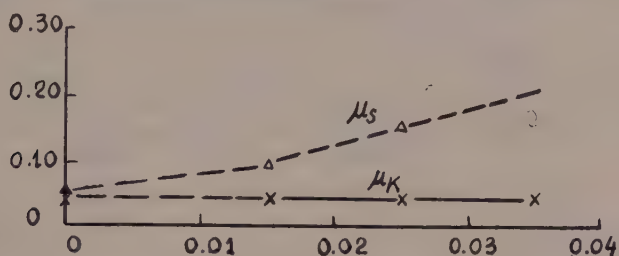


Fig. 4 — Friction of plastics (¹) Phenolic resin.

roughness the situation was similar to the case of low water content of snow (Fig. 5) (¹¹).

In the case of paraffin and skiing waxes (climbing wax and downhill wax), whose thicknesses were kept constant, μ_S was found to depend on the penetration, whereas μ_K remained constant in general (Fig. 6) (¹²).

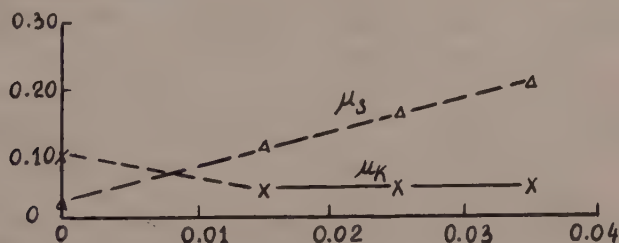


Fig. 5 — Friction of plastics (¹¹) Phenolic resin on slush.

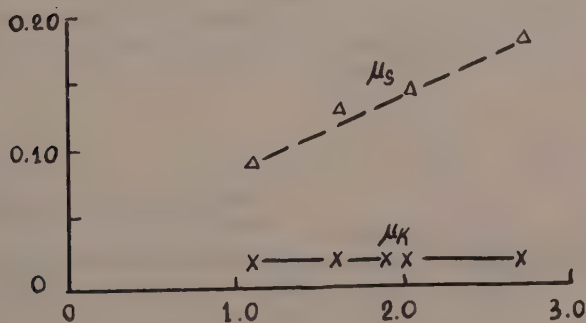


Fig. 6 — Friction of paraffin.

When the penetration was kept constant and the thickness was varied, μ_S increased with increasing thickness while μ_K remained constant (Fig. 7, 8). When, however, the penetration was made larger or the air temperature was made lower, μ_K proved to increase slightly with thickness (Fig. 9) (¹²).

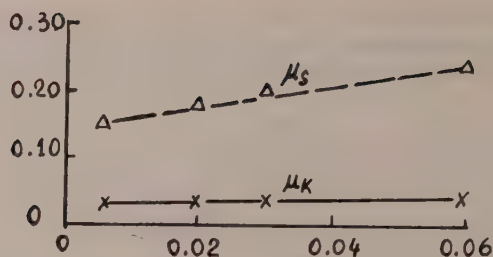


Fig. 7 — Friction of Downhill wax (1) (Swedish wax).

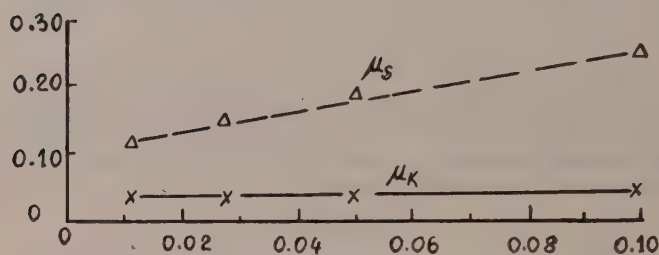


Fig. 8 — Friction of climbing wax (1) (Norwegian wax).

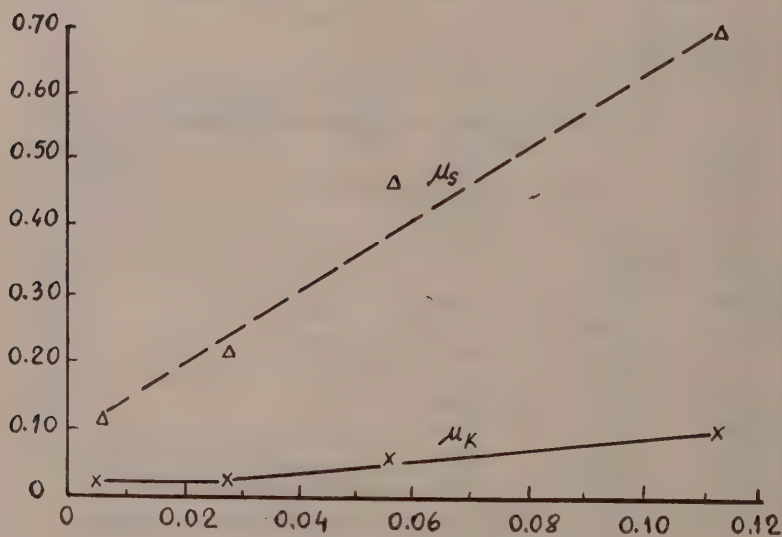


Fig. 9 — Friction of climbing wax (11).

From early days, the behavior of these skiing waxes has been considered to be mysterious, but the above-mentioned data show that climbing ability coincides with the value of μ_s which varies with their penetration, film thickness or other factors and that sliding ability depends upon μ_k on snow.

Thus we can obtain a wax with large μ_S by adjustment of these factors maintaining the value of μ_K constant. The ability of these paraffin and skiing waxes can be explained from their friction-thickness diagram (Fig. 7, 8, 9) ⁽¹²⁾.

The above-mentioned phenomena shown by plastics, paraffin and skiing waxes could be reproduced likewise with crushed ice (Fig. 10). When the water contents of snow and crushed ice were comparable, μ_K had the same value in both cases but μ_S of snow exceeded that of crushed ice (Fig. 10). The reason for this difference is probably to be found in the difference in grain size of snow and crushed ice. Thus the friction of plastics, paraffin, down-hill wax and climbing wax may well be studied in an ordinary laboratory by the use of crushed ice ⁽¹⁴⁾.

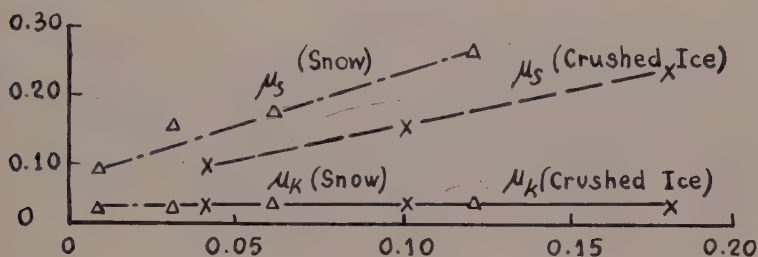


Fig. 10 — Friction of climbing wax on Snow and crushed Ice (Swedish wax).

As regards the defacement of paraffin and skiing waxes on snow, we used radio-isotope and found that the layer of paraffin or skiing waxes was forced by stream of snow to flow from tip to tail with its defacement. The defacement and movement of paraffin or waxes are governed by their rheological properties, and by flexibility of ski ⁽¹³⁾.

Now that much information about the friction of snow was obtained, we finally used organic crystals (diphenyl and maleic acid anhydride) in place of snow and measured friction in the neighbourhood of their melting points ⁽¹⁴⁾.

In this case, both for flat and rough sliding surfaces of phenolic resin, μ_S and μ_K at low temperatures were found to take nearly the same values as in the case of snow and to decrease gradually as the temperature approached the melting point of organic crystal (Fig. 11). Above melting point, flat surface with excess melt showed abnormal increase of μ_K and decrease of μ_S as was the case with snow ^(11, 14).

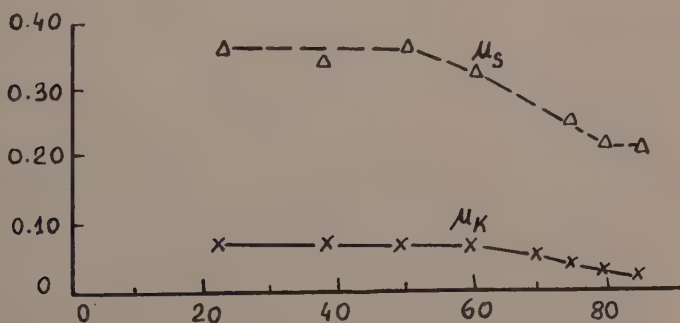


Fig. 11 — Friction on organic crystals (Diphenyl).

From old times, the phenomena due to the friction of snow has been considered as quite peculiar to snow, but the above-mentioned data show that these phenomena are common to all kinds of crystals, organic and inorganic. It is true, in the case of snow crystal, molten crystal (water) absorbed by sliding surface or suspended between sliding surface and compressed snow layer causes some complication, but it has now been shown that the same sort of complication arises also in the frictional phenomena of other kinds of crystals in the neighbourhood of their melting points ⁽¹⁴⁾.

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ROLE OF THE SNOW COVER IN NATURE

G. RICHTER

U R S S

SUMMARY

The snow cover is one of important factors in the development of nature of the moderate and Arctic belts.

Owing to a low thermoconductivity and gas permeability, the snow cover forms an insulating layer, which hinders or stops the heat and gas exchange between the soil and the atmosphere.

Under the protection of the snow both soil and geomorphological processes proceed quite differently than in snowless regions.

The mass of snow protects the soil from a deep freezing and from drastic temperature fluctuations and creates favourable conditions for hibernating plants and animals to survive the hard winter period.

The regime of surface and underground waters is to a great extent determined by the distribution of snow masses and by the character of snow melting.

The snow cover forms a surface sharply differing by its properties from the land surface. This is reflected in the character of all meteorological processes.

The role of the snow cover is not limited by the winter period only; its after-effect is felt quite substantially in the summer as well.

INFLUENCE DU RELIEF SUR L'ACCUMULATION ET LA FONTE DES NEIGES

P. SALAMIN (Hungary)

Université du Bâtiment et des Communications, Budapest

RÉSUMÉ

L'étude suivante résulte en grande partie des mesures effectuées dans les Monts Mátra (dans le bassin versant du ruisseau Kövicses, tableaux 1-2. et figures 1-2.) pendant les mois de février, mars et avril 1956. Toutefois, elle tient compte également des résultats de mesure obtenus, en partie, dans d'autres régions au cours de la période 1954-1960 (tableau 3.) Elle est un exemple intéressant de la recherche méthodique pratiquée sur un terrain de vaste étendue.

Les résultats les plus intéressants, qui se dégagent de nos recherches, sont les suivants (tableau 4-5., figures 4-5.):

1. *Les macroformes de la surface exercent une influence évidente sur l'accumulation et la fonte de la neige.* Mais cette influence ne se manifeste pas toujours très nettement, à cause de la variété des éléments exerçant leur action sur la surface. La quantité de *neige fraîche* peut être des plus variées à n'importe quel endroit de la surface, en fonction de la direction et de la force du vent. L'influence du vent est d'une grande importance surtout dans les hautes plaines, les arêtes et les passes, elle est un peu moins importante dans le fond des vallées. Le manteau de neige *disparaît rapidement* dans le fond des vallées larges, sur les flancs exposés au Sud, dans les plateaux, les arêtes et les pas. A ces endroits, la teneur en eau du manteau de neige est faible, en général, et la densité est élevée. La couche de neige *persiste plus longtemps* sur les flancs orientés vers les O-N-E et, souvent, dans les vallées secondaires étroites aussi. Là, la teneur en eau est généralement élevée, et la densité faible. A un moment donné, la valeur du *changement de la teneur en eau* (Δh) atteint 20 mm. La valeur du *changement de la densité* ($\Delta \gamma$) varie entre 0,01 et 0,02, mais elle peut atteindre également 0,06.

2. *L'influence de l'orientation se manifeste d'une manière plus nette que celle des macroformes.* Sur les superficies situées *vers le Nord*, ou *approximativement vers le Nord*, la teneur en eau est grande et la densité faible, sur celles orientées *vers le Sud* ou *approchant le Sud*, la situation est inverse. Cependant, la différence entre les deux orientations ne représente, numériquement, qu'une influence assez médiocre, si l'on compare le manteau de neige des superficies correspondant au segment de 225° orienté vers les O-S-E avec celui des superficies contenues dans le segment de 135° et situées vers les NO-N-NE. Dans ce cas, la *valeur de la densité varie* entre 0,00 et 0,025 dans la région montagneuse à pente douce et peut s'élever à la valeur de 0,045 dans les montagnes abruptes. Dans le premier cas, la *teneur en eau ne change* que de quelques mm, dans le second, elle peut atteindre 50 mm aussi. Dans les bassins versants *entièrement exposés au Sud*, la fonte des neiges devance souvent de plusieurs semaines la disparition du manteau de neige dans les autres endroits. Ici, l'influence est considérable même numériquement. L'influence de l'orientation se manifeste d'une manière tranchante dans les vallées abruptes longées de l'Ouest à l'Est où les données relatives au manteau de neige couvrant les flancs orientés vers le Sud peuvent être comparées avec les relevées se rapportant aux flancs orientés vers le Nord (Fig. 6.).

3. *L'influence de l'altitude est évidente également.* Avec l'*augmentation de l'altitude*, la teneur en eau *s'élève* et la densité *baisse*. Dans le cas extrême, le *changement de la teneur en eau* est d'environ 40 mm, celui de la *densité* varie, à l'intérieur du bassin versant tout entier, à un moment donné, entre 0,025 et 0,040, mais peut arriver à la valeur de 0,100 également.

4. Dans les parties à *pente douce* des Massifs centraux hongrois, l'influence de l'*altitude* reste la plus importante. Dans les versants composés de parties à *pente abrupte*, l'influence exercée par l'*orientation* constitue l'influence décisive.

5. *Les différents facteurs de relief et la végétation exercent ensemble leur influence.* Cette influence conjuguée peut équilibrer les valeurs des facteurs caractéristiques du manteau de neige, valeurs susceptibles de varier au cours du vieillissement, mais elle peut totaliser aussi les influences concordantes. Dans ce dernier cas, la valeur du *changement de la teneur en eau* peut atteindre, et dépasser même, les 100 mm et celle de la *densité* arriver à la hauteur de 0,20. (Fig. 7.).

* * *

Les résultats exposés ci-dessus semblent présenter un double intérêt : d'une part, ils permettront d'établir à des endroits judicieusement choisis dans les montagnes hongroises, des stations de mesure pour l'observation suivie de la neige; d'autre part, ils fournissent un instrument de recherche commode pour évaluer, sur la base des données relevées par ces stations, le volume d'eau emmagasiné dans les bassins versants de ces montagnes.

1. LE TERRAIN EXAMINÉ, LES MÉTHODES DE RECHERCHE

Le bassin versant du ruisseau *Kövcses* (terrain examiné en premier lieu) couvre une superficie de 44,8 km² et fait partie, selon S. Láng ⁽⁴⁾, des quatre régions morphologiques différentes des Monts *Mátra* : notamment du plateau de *Mátra*, de la région limitrophe du « pied nord du *Mátra* » et du « *Mátra* de *Pásztó* » et, enfin, du « Bassin de *Pásztó* » (Fig. 1-2).

Le Plateau de *Mátra* coïncide à peu près complètement avec le bassin versant supérieur du ruisseau *Kövcses*. Il est une *pénéplaine faiblement ondulée, aplanie* par les érosions, exception faite pour quelques monts coniques sur les bords (*Ágassvár*, *Óvár*, *Tóthegyes*). Les vallées principales (vallées du ruisseau *Kövcses* et du ruisseau *Csörgő*, son plus grand affluent) en s'éloignant de la région des sources vers l'Est, s'enfoncent profondément dans le terrain et le découpent de plus en plus fortement.

Le prolongement du pied nord du *Mátra* s'étendant le plus loin vers l'Ouest, constitue la plateau de *Gombás* et borde au Nord le tronçon inférieur de la vallée du *Kövcses*. Haut de 406 mètres, le plateau de *Gombás* est une *pénéplaine aplanie* assez surélevée formant un plan horizontal. Sur la surface aplanie par les érosions, un cône de débris s'est formé.

La partie du *Mátra* de *Pásztó* la plus avancée vers le Nord, le mont *Nyikom*, haut de 760 mètres, constitue la limite sud du tronçon inférieur de la vallée du *Kövcses*. Le mont *Nyikom* descend vers la vallée en pente rapide, hérissée par endroits d'escarpements à pic. Le *Mátra* de *Pásztó* et le *Nyikom* surplombent le plateau de *Gombás* aligné au pied nord du *Mátra* et forment ainsi une vallée asymétrique avec le lit du *Kövcses*.

A l'extrémité du tronçon inférieur de la vallée du *Kövcses* faisant partie du Bassin de *Pásztó*, la vallée s'est creusée dans la surface d'un cône de déjection de grande étendue qui sert de transition entre la région de collines et le plateau.

Pour représenter le caractère général du relief de la topographie, nous avons reproduit dans les Figures 1-2, la répartition proportionnelle des superficies par rapport à l'altitude et à la pente.

Le point le plus bas (dans le Bassin de *Pásztó*) s'élève à 175 mètres et le point le plus haut (le plateau de *Piszkés*) à 946 mètres au-dessus de la mer Adriatique. Le climat du bassin versant a été analysé dans deux études de M. Kéri ^[1-2].

Aux niveaux inférieurs, le bassin versant est revêtu de forêts de chênes sur les plateaux les plus élevés et sur les flancs exposés au Nord, de forêts de hêtres. En beaucoup d'endroits, on relève des prêtres et des pâturages. Des champs d'une certaine étendue n'apparaissent que dans les parties les plus basses, à l'Ouest (dans le Bassin de *Pásztó*).

La terre est un sol forestier, sombre sur les parties supérieures du plateau, décolorée au milieu et brune dans les parties basses.

Pour mener des recherches parallèles, nous avons choisi la partie supérieure ($F = 16,7$ km²) du bassin versant du ruisseau *Garadna* décrite dans nos études précédentes ^[7-9]. Ici, on s'est contenté d'esquisser, à côté des caractéristiques du bassin versant du *Kövcses*, la répartition de la superficie par rapport à l'altitude et à la pente (Fig. 1-2). Il est à remarquer que les pentes sont beaucoup plus douces dans le bassin versant du *Kövcses* que dans celui de la *Garadna*.

Le bassin versant du *Kövcses* est caractéristique, au point de vue de la fonte des neiges, pour la plus grande partie des Massifs centraux hongrois descendant en pentes douces, tandis que le bassin versant de la *Garadna* rappelle quelques escarpements plus accentués de ces massifs.

Dans nos études précédentes, nous avons exposé les principes et les méthodes de nos recherches qui avaient été appliqués dans le choix des emplacements pour les prélèvements d'échantillons ^[3,7-9].

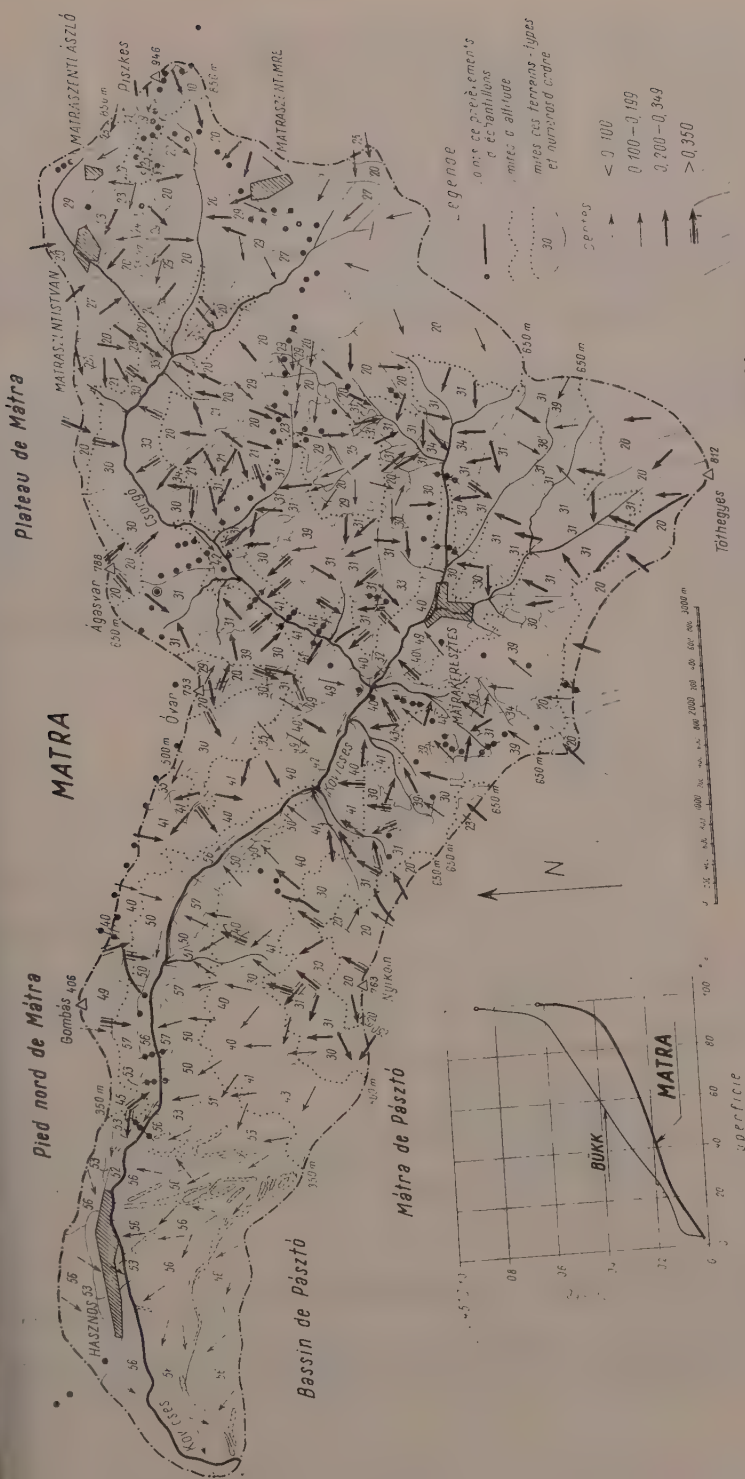


Fig. 2 — Les pentes du bassin versant du ruisseau Kővisek. Monts Mátra.

Dans l'élaboration des méthodes de recherches, nous nous sommes efforcés de tenir compte de tous les phénomènes de l'économie thermiques et de l'économie d'eau. (Parmi les premiers, nous avons considéré comme les plus importants : la condensation ou la transformation en état liquide de la teneur en vapeur d'eau contenue dans les interstices et sur les surfaces du manteau de neige; la radiation solaire; la radiation de l'air dans la direction de la couche de neige; l'échange de chaleur entre l'air et la couche de neige; l'effet de la pluie tombée sur cette dernière et, enfin, la conduction de la chaleur à partir de l'air et du sol. Dans le domaine de l'économie d'eau, nous avons examiné comme des facteurs essentiels : le rôle des précipitations sous forme de pluie et de neige; le rôle de la teneur en vapeur d'eau se transformant en état liquide ou solide à l'intérieur ou à la surface du manteau de neige et, enfin, l'évaporation, avec le ruissellement, l'écoulement et l'infiltration de la neige fondue [1-9]).

Au cours de nos recherches, nous avons déterminé tout d'abord l'épaisseur du manteau de neige, sa teneur en eau, sa densité, et délimité les superficies-types ayant un caractère identique au point de vue de l'accumulation et de la fusion de la neige. Pour le bassin versant du *Mátra*, la répartition de ces superficies-types est représentée dans les Figures 1-2. (En rapport avec les recherches effectuées dans les Monts *Bükk*, nous avons publié déjà l'analyse du terrain examiné [9]). Dans la Figure 1, nous reproduisons l'état de la culture du terrain et le numéro d'ordre des parties de superficies délimitées. Sur la Figure 2, nous indiquons les points cardinaux, les pentes, ainsi que les numéros-types des superficies caractéristiques.

Le numéro-type se compose de deux chiffres dont le premier indique l'altitude selon l'échelle ci-dessous :

- | | |
|---------|--|
| 1. | > 850 m au-dessus de la mer Adriatique |
| 2. | 650-850 m » » » » » |
| 3. | 500-650 m » » » » » |
| 4. | 350-500 m » » » » » |
| 5. | < 350 m » » » » » |

le deuxième chiffre caractérise les rapports du relief et la végétation :

- | | |
|-----------|---|
| 0-1 | surface boisée, notamment : |
| 0 | flanc de large vallée principale, crête et fond de vallée correspondante |
| 1 | vallée secondaire étroite |
| 2-5 | terrain planté de jeunes arbustes, terrains broussailleux, bosquet, notamment : |
| 2 | fond de vallée |
| 3 | flanc de vallée principale |
| 4 | vallée secondaire |
| 5 | plateau, crête, pas |
| 6-9 | pré, pâturage, champ, notamment : |
| 6 | fond de vallée |
| 7 | flanc de vallée principale |
| 8 | vallée secondaire |
| 9 | plateau, crête, pas. |

(Ainsi, la surface 21 désigne une vallée secondaire étroite et boisée située à une altitude de 650 à 850 mètres.)

Afin de caractériser la nature des 156 emplacements choisis pour les prélèvements d'échantillons, nous présentons, dans le Tableau 1, la répartition de ces emplacements selon les superficies-types.

TABLEAU 1

Répartition des points de mesures de la bassin versant du ruisseau Kővicses

Caractère des emplacements pour les prélèvements d'échantillons	Nombre des points de mesures à l'altitude de					au total
	> 850	850-650	650-500	500-350	< 350	
	au-dessus de la mer Adriatique					
fond de vallée	—	3	—	3	14	20
vallée secondaire	—	—	16	13	—	29
flanc	1	18	19	10	—	48
plateau	13	25	11	10	—	59
au total	14	46	46	36	14	156
O-S-E	5	21	20	8	5	59
NO-N-NE	4	14	17	15	3	53
surface horizontale	5	11	9	13	6	44
au total	14	46	46	36	14	156
forêts	1	7	26	19	—	53
terrains broussailleux	7	7	10	10	4	38
pâturages-prés-champs	6	32	10	7	10	65
au total	14	46	46	36	14	156

Le changement des conditions météorologiques les plus importantes survenues durant la période décisive des recherches est détaillé dans la Figure 3, établie sur la base des observations faites quotidiennement par la station météorologique de Mátraszentlászló.

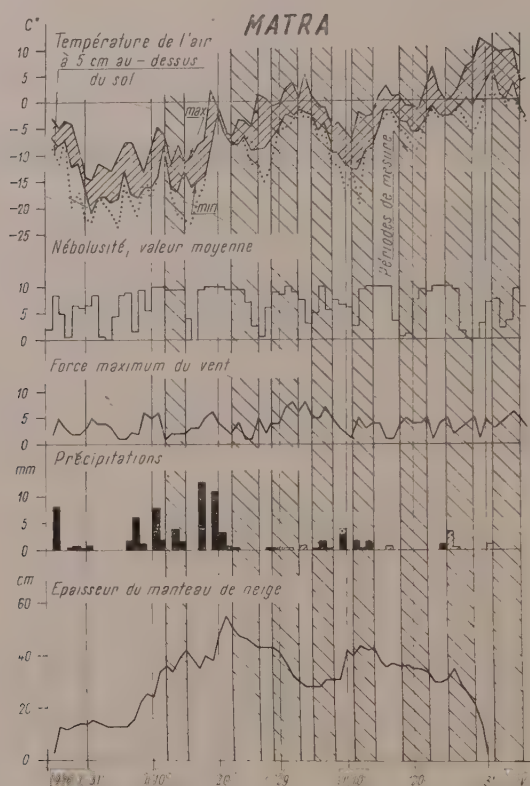


Fig. 3 — Le changement des conditions météorologiques. Monts Mátra, 26.1.-6.4. 1956.

La variété des champs d'expériences et la diversité des emplacements choisis pour les prélèvements d'échantillons nous a permis [9], d'une part, de reconnaître le caractère régulier des faits valables sur une superficie de grande étendue et, d'autre part, d'étudier séparément tous les facteurs influant sur l'accumulation et la fonte des neiges.

2. LES ÉTAPES DE LA RECHERCHE. LES RÉSULTATS

L'étude des facteurs de relief influant sur l'accumulation et la fusion de la neige a été effectuée dans l'ordre ci-dessous :

TABLEAU 2 (Suite du Tableau 2)

Données hydrologiques du manteau de neige (Mátra, février-avril 1956)

Caractère de la surface	Périodes de mesures						
	13-15.II	23-26.II	29-13.III	6-8.III	12-14.III	19-22.III	26-29.III
	Valeur de la teneur en eau h mm						
forêts	32,2	66,4	65,0	63,3	70,8	64,2	59,0
terrains broussailleux	33,4	67,9	71,1	73,5	80,4	76,5	63,5
pâturages-près-champs	34,1	65,9	68,3	70,6	77,7	77,9	66,3
la moyenne	33,2	66,5	67,7	68,5	75,7	72,3	63,1

Valeur de la densité γ kg/l

fond de vallée	0,132	0,260	0,317	0,326	0,285	0,346	0,328	—
vallée secondaire	0,135	0,217	0,260	0,318	0,292	0,343	0,345	0,384
flanc	0,132	0,204	0,251	0,291	0,280	0,327	0,318	0,406
plateau	0,143	0,220	0,269	0,309	0,286	0,322	0,333	0,408

Caractère de la surface	Périodes de mesures						
	13-15.II	23-26.II	29-13.III	6-8.III	12-14.III	19-22.III	26-29.III
	Valeur de la densité γ kg/l						
O-S-E	0,131	0,221	0,267	0,306	0,290	0,336	0,326
NO-N-NE	0,137	0,212	0,250	0,299	0,266	0,321	0,325
au-dessous de 350 m	0,134	0,292	0,385	0,379	—	—	0,404
entre 350 et 500 m	0,134	0,234	0,278	0,331	0,296	0,359	—
entre 500 et 650 m	0,136	0,209	0,254	0,296	0,292	0,334	(0,394)
entre 650 et 850 m	0,139	0,210	0,260	0,294	0,273	0,300	0,375
au-dessus de 850 m		0,195	0,232	0,287	0,275	0,303	0,413
forêts							0,408
terrains broussailleux	0,133	0,213	0,263	0,311	0,290	0,338	0,340
pâturages-prés-champs	0,133	0,219	0,260	0,287	0,276	0,330	0,401
	0,140	0,226	0,275	0,315	0,286	0,322	0,370
la moyenne							0,429
	0,136	0,220	0,266	0,307	0,286	0,329	0,330
							0,405

TABLEAU 3

Données hydrologiques du manteau de neige (Bükk, février 1956)

Caractère de la surface	Valeur de la teneur en eau h , mm		Valeur de la densité γ , kg/l	
	12-15. II	26-28. II	12-15. II	26-28. II
fond de vallée	72,0	83,0	0,180	0,220
vallée secondaire	73,0	88,0	0,160	0,220
flanc	66,7	73,3	0,163	0,226
plateau	73,2	90,0	0,175	0,239
S	56,0	20,0	0,215	0,293
O-SO et SE-E	73,5	68,0	0,155	0,230
O-SO-S-SE-E	61,8	34,0	0,195	0,275
N-NO-NE	68,7	85,7	0,150	0,210
au-dessous de 600 m	65,4	72,8	0,174	0,228
entre 600 et 750 m	69,3	78,4	0,164	0,229
au-dessus de 750 m	75,8	95,1	0,165	0,229
vieilles forêts ⁺	73,0	83,0	0,169	0,223
forêts ⁺	63,5	79,2	0,154	0,218
jeunes forêts ⁺	70,2	85,2	0,166	0,224
prés-pâturages	73,5	92,8	0,165	0,241
la moyenne	69,7	80,2	0,168	0,229

Remarques : + Forêts d'arbres à feuilles caduques.

I. — Les macroformes de la superficie (les parties inférieures et les flancs de vallées principales larges et étroites, les vallées secondaires étroites, crêtes, pas, plateaux etc).

II. — Orientation : vallées, flancs de vallée situés vers le N, le NE, l'E etc.

III. — Position par rapport à l'altitude (superficies au-dessus de 850 m; superficies situées entre 850 et 650 m, 650 m et 500 m, et 500 et 350 m d'altitude; superficies au-dessous de 350 m d'altitude).

Au cours des travaux de recherches, on a examiné séparément et étudié dans leur ensemble aussi l'influence des facteurs de relief, en premier lieu dans les cadres de la figure morphologique du bassin versant du Kővicses, et en second lieu, dans ceux de la Garadna.

3. L'INFLUENCE DES MACROFORMES DE LA SUPERFICIE SUR L'ACCUMULATION ET LA FUSION DE LA NEIGE

Nous allons examiner maintenant tour à tour le rôle que jouent dans l'accumulation et la fusion de la neige, d'abord la partie inférieure (1) et les flancs (3) des vallées principales larges et étroites, puis l'ensemble des vallées secondaires étroites (2) et, enfin, les crêtes, les pas et les plateaux (4). (Tableaux 2-3, Figures 4-5 [7⁸]).

3.1. L'influence des fonds de vallées principales

La quantité de *neige fraîche* tombée sur la partie inférieure des vallées, *le fond de vallée*, dépend principalement du fait que la chute de neige a lieu par temps venteux ou quand il ne fait pas du vent. *S'il vente*, la quantité de neige parvenue au fond de vallée peut être plus ou moins importante que celle tombée sur les terrains ouverts, exposés au vent (flancs et plateaux), suivant la direction et la force du vent, l'orientation de la vallée et la place qu'elle occupe dans l'espace. *Par temps calme*, il tombe plus de neige sur le fond de vallée approximativement horizontal que sur les flancs déclinés. Cependant, l'excédent ne doit pas être trop considérable. La neige parvenue déjà sur le fond de vallée est plus à l'abri du vent que celle tombée sur des terrains plus découverts.

La quantité de neige tombant sur le fond de vallée est déterminée également par *la végétation et la position relativement à l'altitude*. La partie inférieure des larges vallées principales, généralement, n'est pas revêtue de bois, il n'y a que de bosquets; des buissons et de petits bois ne se trouvent qu'à la lisière. De ce fait, elle peut retenir plus de neige fraîche que d'autres terrains [9]. Toutefois, le fond de vallée profondément enfoncé fait valoir déjà l'influence de l'altitude (qu'on analysera tout à l'heure), les terrains enfoncés recueillant moins de neige. Mais, dans le cas donné, l'influence est médiocre, puisque ce n'est pas la profondeur du fond de vallée qui détermine, au point de vue de l'altitude, la quantité de la neige tombée, mais l'altitude de la partie de montagne où la vallée s'est creusée sa place.

Donc, la quantité de neige fraîche tombant sur le fond de vallée *peut être plus ou moins importante* que la teneur en eau du manteau de neige formé sur les terrains ouverts.

Au cours de la *condensation* et de la *fusion* de la neige, appelées aussi *vieillessement*, le caractère de la teneur en eau du manteau de neige et, d'une manière parallèle, le caractère de sa densité peuvent être déterminés déjà avec plus de précision. En général le manteau de neige a *une teneur en eau (*) plus faible et une densité plus forte sur le fond de vallée* que sur les flancs et les plateaux. Dans les vallées favorablement

(*) Par l'expression «teneur en eau» nous comprenons l'eau sous forme solide et liquide également. Ainsi, il aurait été plus logique d'employer ici le terme technique un peu lourd mais plus précis de l'*«équivalent d'eau»*.

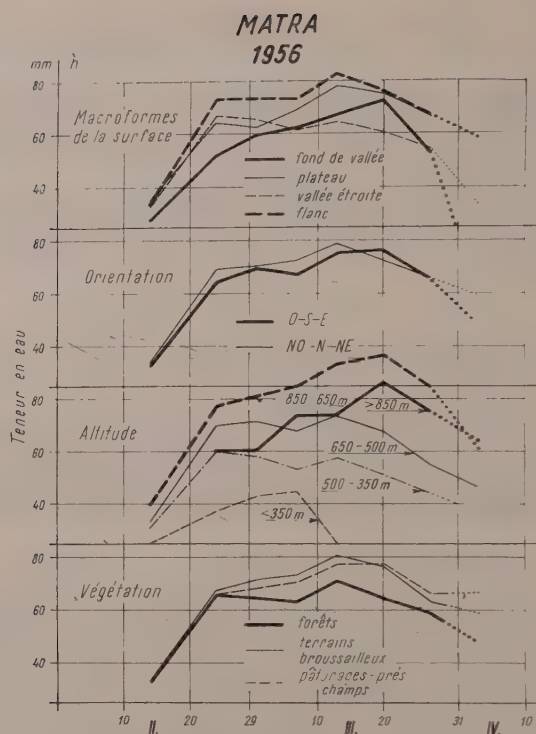


Fig. 4 — L'influence des facteurs de relief sur la teneur en eau. Monts Mátra.

exposées par rapport aux points cardinaux, la radiation solaire, la conduction de la chaleur à partir du sol et les autres phénomènes thermiques peuvent exercer leur plein effet, en conséquence le manteau de neige fond plus vite, sa teneur en eau diminue, sa densité augmente. Un cas peut constituer une *exception*, c'est quand au cours de l'amoncellement de la neige une quantité plus considérable de neige s'entasse au fond de vallée que sur les flancs et les plateaux. Or, même dans ce cas, avec la progression de la fusion, la teneur en eau de la couche de neige étendue sur le fond de vallée diminuera de plus en plus et descendra au-dessous de celle de la neige couvrant les flancs et les plateaux tandis que sa densité changera en sens inverse. La règle comporte encore une autre exception. Dans le cas des vallées étroites, quand l'économie d'énergie thermique du manteau de neige est défavorable pour la fusion, et la radiation solaire p.e. ne peut exercer ses effets, la neige se conserve là plus longtemps que sur les flancs ouverts et les plateaux.

Ces remarques théoriques sont vérifiées par les *résultats d'essais*.

En 1956, dans les champs d'expériences des *Mátra*, on a observé que la teneur en eau du manteau de neige était généralement plus faible dans les parties profondes des vallées que sur les terrains ouverts, conformément à la progression de la fusion dans une zone abritée mais non complètement protégée. Alors que la densité de la neige était plus élevée à cet endroit que partout ailleurs (Figure 5a). Cette divergence a été la plus sensible pendant la période du 23 février au 8 mars quand la neige persistait encore dans la vallée du ruisseau Kövicses, dans le Bassin de Pásztó, tandis que la

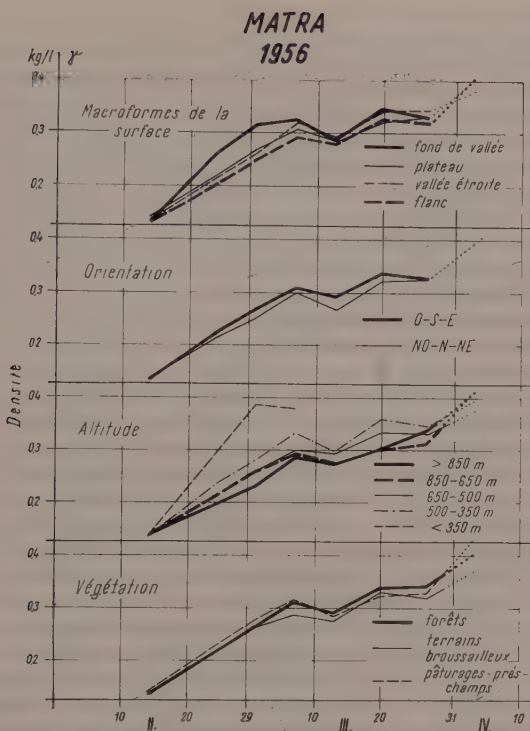


Fig. 5 — L'influence des facteurs de relief sur la densité. Monts Mátra.

fusion était déjà très avancée sur le large fond de vallée à une altitude de 250 à 300 m et la densité atteignait une valeur relativement élevée.

Dans les champs d'expériences des Monts *Bükk* (Tableau 3), en 1954, la teneur en eau du manteau de neige, dans la partie inférieure beaucoup plus étroite de la vallée principale, dépassait la moyenne et la valeur de la densité variait autour de celle-ci ou la dépassait également.

Ainsi, les résultats enregistrés dans les Monts *Mátra* confirment la conclusion générale, tandis que ceux de la région étudiée dans les *Bükk* donnent l'exemple de l'exception.

3.2. L'influence des fonds de vallées secondaires

Les remarques formulées au sujet de la vie de l'enneigement persistant dans les fonds de vallée plus larges sont valables également pour l'accumulation et le vieillissement du manteau de neige dans les vallées secondaires étroites (le fond de vallée et les flancs considérés tout ensemble). Cependant, l'état de leur encaissement exerce une influence protectrice sur le changement des propriétés physiques de l'enneigement et le rend plus équilibré. Dans les Monts *Mátra*, massif moins abrupte, la teneur en eau, en général, était inférieure (Figure 4a) et la densité supérieure à la valeur moyenne (Figure 5a). Mais, ici l'écart était moins important que pour l'enneigement des fonds de vallées larges. Dans le bassin versant des Monts *Bükk*, versant en pente plus rapide [8], la teneur en eau a dépassé la moyenne, mais la densité restait au-dessous de celle-ci, ainsi que nous l'avons remarqué plus haut.

3.3. L'influence des flancs

La quantité de la *neige fraîche* tombant sur les flancs est fonction de la force et de la direction du vent. Par *temps calme*, le manteau de neige a la même épaisseur dans toutes les parties du bassin versant. *S'il fait du vent*, sa teneur en eau est déterminée par le mouvement des masses d'air et le rapport entre les formes de relief dans l'espace. Dans le cas d'un *vent égal mais non pas trop fort*, la couche de neige sera sensiblement différente dans les endroits abrités du vent et sur les flancs libres non protégés. C'est dans de telles conditions qu'on voit se former des bandes de neige plus ou moins importantes, déposées par le tourbillon de neige mais, dans le bassin versant tout entier, considérant tout ensemble les flancs différemment orientés, on obtient des valeurs approchant la moyenne. Si *le vent est très fort*, la situation est toute autre. Sur la plus grande partie des flancs de montagne, il reste à peine de la neige, un enneigement uniforme ne s'accumule que sur les flancs tout à fait protégés du vent. On peut observer des bandes de neige partout. Dans les montagnes de Buda, nous en avons fait l'expérience en janvier 1959, il n'y avait de manteau de neige continu qu'à quelques endroits à l'abri du vent et la teneur en eau était des plus élevées.

En considérant, dans le même esprit, les flancs différemment orientés pendant la période du vieillissement et de la fonte du manteau de neige, on remarquera que la fusion s'accomplit lentement, la teneur en eau de l'enneigement est généralement importante, et même très élevée, et la densité moyenne reste faible. Ce phénomène est la conséquence naturelle du fait que sur les trois quarts de la circonférence décrite par le cercle de 360 degrés, à l'exception des flancs exposés au Sud, au Sud-Est et au Sud-Ouest, la *radiation solaire* exerce ses effets pendant un espace de temps plus restreint que sur les fond de vallée approximativement plats et les hautes plaines respectivement. En réunissant tous les relevés d'observation relatifs aux flancs de montagne, on arrive nécessairement à cette conclusion. Elle se fait remarquer surtout, comme nous l'avons déjà dit, quand on retranche de l'examen les flancs exposés au Sud. D'ailleurs, ces propositions s'appuient sur l'*influence de la végétation* aussi [9]. Les flancs de montagnes sont généralement couverts de forêts, de ce fait la teneur en eau est relativement faible au moment de la formation du manteau de neige fraîche mais, au cours du vieillissement des neiges, la végétation peut déjà exercer une action préservatrice et retarder le développement de la fusion, donc la densité restera plus faible et la teneur en eau sera plus forte que la valeur moyenne.

Quelques *relevés d'observation* peuvent montrer l'évidence de ces propositions. Les résultats des mesures effectuées dans les *Mátra* (Tableau 2 et Figures 4a-5a) et dans les *Bükk* (Tableau 3) sont parfaitement concordants. (Pour les recherches dans les *Bükk*, les données en rapport avec les flancs exposés au Sud sont présentées séparément).

3.4. L'influence des crêtes de montagne, des pas et des plateaux

Sur les crêtes de montagne, les pas et les plateaux, l'accumulation et la fonte de la neige ne s'accomplit pas de la même manière que sur les fonds de vallée et les flancs.

Dans l'accumulation de la *neige fraîche*, le facteur le plus important est, là aussi, le *vent* exerçant librement ses effets extrêmes dans les aires de surélévation. Le vent peut trans former à sa guise la couche de neige fraîche peu consistante. Les bandes de neige constituées par les tourbillons de neige sont nombreux et on voit des superficies préservées de la neige à peu près complètement. (Dans les *Mátra*, on a observé ces faits, en particulier, sur la haute plaine s'étendant au-dessous du plateau de Piszksés) On reconnaît clairement les effets de l'*altitude* et de la *végétation* aussi. Dans ces zones, naturellement surélevées conformément à leur caractère, l'influence de l'altitude qu'on va étudier tout à l'heure, apparaît déjà dans toute son ampleur. La végétation

habituelle de la haute plaine contribue également à donner une forme caractéristique à l'accumulation et à la fonte des neiges. Les hautes plaines se composent, dans la majorité des cas, de superficies couvertes de bosquets ou de champs et de pâturages entourés de terrains broussailleux et boisés. (Ces terrains retiennent beaucoup plus de neige [9]). Compte tenu de tous ces facteurs, on peut *énoncer qu'en général*, sur les terrains de haute plaine, sans végétation et librement exposés au vent, la teneur en eau de la couche de neige fraîche est faible tandis que sur les plateaux revêtus de broussailles et sur les champs, pâturages et près des plateaux élevés, encadrés de bois ou de terrains broussailleux la teneur moyenne en eau du manteau de neige dépasse de peu ou de beaucoup la valeur moyenne relevée dans le bassin versant tout entier. Cependant, remarquons qu'il est assez difficile de déterminer la situation générale exacte, par suite de la grande variété que présente le manteau de neige. Mais, si les emplacements destinés aux prélèvements d'échantillons sont judicieusement choisis, on arrive tout de même à observer les principaux changements successifs de la vie de la neige.

Dans la couche de neige couvrant le haut plateau et dont la *condensation* et la *fusion* est déjà en marche, la teneur en eau peut être quelquefois plus faible que la moyenne mais, en général, elle la dépasse conformément aux influences variées dont nous avons parlé plus haut. Durant la première partie de la période de condensation et de fusion, la teneur en eau du manteau de neige, dans la majorité des cas, dépasse la valeur moyenne, ce qui correspond au caractère naturel de ces terrains broussailleux. Mais, dans la dernière phase de la fusion, la teneur en eau peut tomber au-dessous de la moyenne. La densité est presque toujours au-dessus de la moyenne. Sa valeur élevée est le résultat du fait que sur les hauts plateaux la radiation solaire de même que les autres phénomènes thermiques (l'échange de chaleur entre la couche de neige et l'air, la conduction de la chaleur à partir de l'air et du sol) peuvent exercer librement leurs effets si les conditions météorologiques sont favorables, et la condensation et la fusion s'accomplissent ainsi plus rapidement.

Les résultats des recherches poursuivies dans les *Mátra* (Tableau 2, Figures 4a-5a) et dans les *Bükk* (Tableau 3) ont confirmé le raisonnement sur tous les points.

Disons, pour conclure, que les macroformes de la surface exercent, en tout état de cause, une influence évidente sur l'accumulation et la fonte de la neige. Cette influence se manifeste quelquefois sous des formes contradictoires. Elle se présente, à maintes reprises, comme une force très importante, à la suite d'autres influences agissant dans le même sens (influences des points cardinaux, de l'altitude et de la végétation). Prenons par exemple les Massifs centraux hongrois : leurs flancs sont généralement boisés, les plateaux couverts de bosquets, les fonds de vallée, par leur caractère même, profonds et les plateaux élevés etc. Dans ces conditions, il est difficile de démontrer l'influence immédiate du relief, mais elle existe pourtant et se révèle souvent considérable.

4. L'INFLUENCE DE L'ORIENTATION DU BASSIN VERSANT SUR L'ACCUMULATION ET LA FUSION

La divergence entre l'orientation vers le Nord ou le Nord-Est, etc. est une chose allant de soi et découle de la différence entre les économies d'énergie thermique des manteaux de neige exposés à des points cardinaux différents. *Sur les superficies orientées vers le Nord ou approximativement vers le Nord*, la teneur en eau est forte et la densité est faible, sur les versants *exposés au Sud* ou situés approximativement dans cette direction, la situation est inverse. Sur les flancs *exposés entièrement au Sud*, la fonte de neige s'achève souvent plusieurs semaines plus tôt que sur n'importe quel autre terrain. (Dans ce qui précède, il était déjà question de l'effet de la position relativement aux différents points cardinaux mais nous n'avons pas fait une distinction

sur les résultats auxquels avaient abouti les mesures par prélèvements d'échantillons effectuées à des endroits différemment exposés).

En partant des résultats obtenus, on peut établir le fait que dans les Monts *Mátra* à pente plus douce et ayant, en général, le caractère d'un plateau (Tableau 2, Figures 4b-5b), les différences paraissent moins importantes que dans les *Bükk* (Tableau 3) et, en particulier, dans la vallée fortement escarpée du ruisseau Garadna, allongée exactement de l'Ouest à l'Est. Les *Mátra* plus équilibrés au point de vue du relief, le sont aussi par rapport à la teneur en eau et à la densité du manteau de neige. Dans les *Bükk*, par contre, les flancs de montagnes exposés au Sud et au Nord nous ont permis de comparer des flancs exposés en sens inverse. La différence la plus importante entre les flancs situés vers le S et vers les NO-N-NE — pendant la période où le manteau de neige durait encore sur le flanc sud aussi et permettait d'y effectuer des mesures —, a été donnée par les équations $\Delta\gamma = 0,08-0,09$ kg/l et $\Delta h = 75$ mm. Les superficies orientées vers les O-S-E et correspondant à une cercle de 225 degrés d'ouverture, sont, en majeure partie, des superficies ensoleillées pendant le jour et ne diffèrent déjà que peu du manteau de neige des superficies restées à l'ombre pendant la plus grande partie du jour, et correspondant à une ouverture de 135° degrés. Naturellement, la différence apparaît, dans ce cas aussi, très nettement.

Dans la Figure 6, nous reproduisons l'influence de l'orientation sur la densité et la teneur du manteau de neige sur la base des résultats des mesures effectuées dans les Monts *Mátra* sur quelques points caractéristiques.

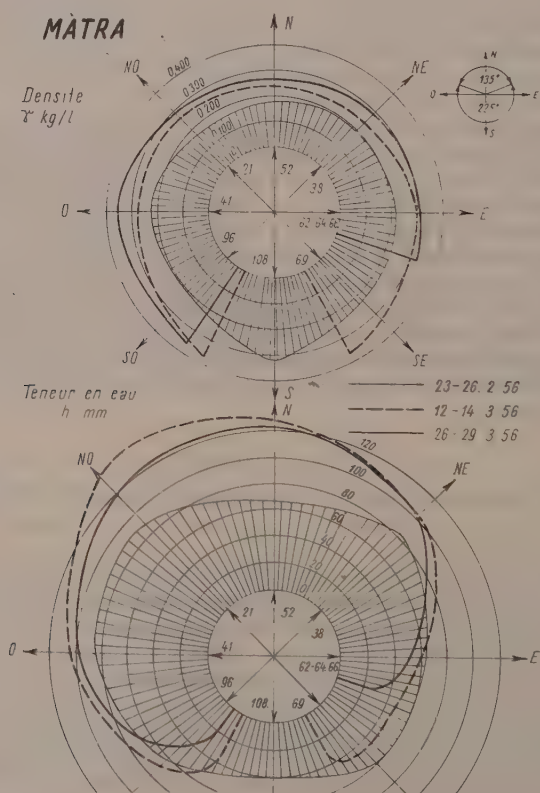


Fig. 6 — L'influence de l'orientation sur l'accumulation et la fusion. (Les numéros 21, etc. indiquent les points de mesures).

5. L'INFLUENCE DE L'ALTITUDE SUR L'ACCUMULATION ET LA FUSION

Il nous reste encore à examiner le rôle de l'altitude. L'influence de l'altitude est évidente : la teneur en eau augmente (Tableau 2-3, Figures 40-50) et la densité diminue avec l'altitude. La température moyenne diminue avec l'accroissement de l'altitude, dans le haut du bassin la fonte des neiges commence plus tard, la condensation et la disparition du manteau de neige s'effectuent plus lentement. Cependant, dans les terrains plus élevés la neige s'accumule en plus grande quantité que dans les superficies basses [7, 8, 9].

Les résultats d'observation recueillis dans les *Mátra* ont confirmé ces règles. Nous n'avons relevé quelque *exceptions* en rapport avec la valeur de la densité qu'au cours de la première et la dernière semaine de nos recherches, dans les terrains d'au-dessus de 850 m d'altitude. Ce qui est dans l'ordre des choses puisque durant la première et la dernière semaine, c'est-à-dire dans les cas de la neige fraîche et de la neige complètement condensée, la densité approche de près la valeur limite inférieure ($\gamma_a = 0,10$) et supérieure ($\gamma_f = 0,36-0,40$) respectivement; à ce point, les données de la densité se trouvent en répartition normale comme nous l'avons indiqué dans nos recherches antérieures [2], donc il n'y a pas de changement décisif. D'autre part, les superficies d'au-dessus de 850 m d'altitude ne représentent qu'une très faible partie du bassin versant total et se composent principalement de terrains exposés au vent, à la radiation solaire et aux autres influences thermiques, si bien que le manteau de neige ne pourra se former qu'inégalement et ne devra avoir qu'une épaisseur relativement faible où la condensation et la fusion se réaliseront rapidement. Ainsi, ces terrains surélevés ne peuvent être considérés comme étant caractéristiques pour toute la superficie de la même altitude parce que cette superficie se compose de terrains différemment orientés qui sont revêtus de végétations diverses et sont différents de relief aussi.

Les résultats de mesures relevés dans les *Monts Bükk* ont confirmé également la règle dont nous avons fait état [8]. A une altitude moyenne de 475-850 m p.e. la teneur en eau de la couche de neige fraîche variait de 93% à 111%. Dans la couche de neige vieillie, le changement était beaucoup plus considérable. La valeur de la densité variait dans la même mesure pour le manteau de neige nouvellement formé aussi bien que pour la neige vieillie. A l'exception, toutefois, des flancs exposés au Nord, protégés de la radiation solaire directe. Pour les flancs, le changement de la densité était le même aux différentes altitudes.

6. L'INFLUENCE DE L'ENSEMBLE DES FACTEURS DE RELIEF SUR L'ACCUMULATION ET LA FONTE DES NEIGES

Nous allons résumer maintenant, sur la base de nos recherches antérieures et compte tenu principalement des résultats d'essai relevés dans les *Monts Mátra*, l'effet numérique de l'ensemble des facteurs de relief dans un bassin versant dont l'altitude varie entre 175-950 m (Tableau 4). Dans le cas donné, la densité moyenne est $\gamma = 0,22, 0,29$ et $0,33$ respectivement. Dans le Tableau 4, nous avons reproduit pour permettre la comparaison, les influences de la végétation antérieurement étudiées aussi [9]. On remarquera que dans les *Monts Mátra*, de même que dans la plus grande partie des Massifs centraux hongrois caractérisés par des conditions de relief semblables, l'altitude représente l'influence la plus considérable. Les macroformes de la surface et la végétation (Tableau 2-3, Figures 4d-5d) doivent exercer une influence à peu près identique sur le développement de la densité alors que la formation de la teneur en eau dépend plus des macroformes de la surface que de la végétation. (Notons que dans ces considérations nous n'avons pas tenu compte des sapinières ne faisant que

TABLEAU 4

Influence des facteurs de relief sur la densité et la teneur en eau du manteau de neige (Mátra, 1956).

Facteurs de relief	Valeur de la variation de la teneur en eau : $\Delta h = h_{max} - h_{min}$ mm si la densité moyenne dans le bassin versant total est			Remarques
	0,220	0,286	0,330	
	et la teneur moyenne en eau dans dans le bassin versant total est			
	66,5	75,5	62,5	
macroformes de la surface	9 (20)	18	14	(00) les fonds de vallée y compris
orientation	6	4	0	les surfaces O-S-E comparées à cel- les NO-N-NE
altitude	18 (40)	35	40	(00) les surfaces au- dessous de 350 m y comprises
végétation	4	10	7	

Facteurs de relief	Valeur de la variation de la densité: $\Delta \gamma = \gamma_{max} - \gamma_{min}$ kg/l, si la densité moyenne dans le bassin versant total est			Remarques
	0,220	0,286	0,330	
macroformes de la surface	0,02 (0,06)	0,01	0,02	(00) les fonds de vallée y compris
orientation	0,01	0,025	0,00	les surfaces O-S-E comparées à cel- les NO-N-NE
altitude	0,04 (0,10)	0,025	0,035	(00) les surfaces au-dessous de 350 m y compri- ses
végétation	0,01	0,015	0,02	

TABLEAU 5

*Influence des facteurs de relief sur la densité et la teneur en eau du manteau de neige
Bükk, 1954).*

Facteurs de relief	Valeur de la variation de la teneur en eau : $\Delta h = h_{max} - h_{min}$ mm si la densité moyenne dans le bassin versant total est		Remarques
	0,170	0,230	
	et la teneur moyenne en eau dans le bassin versant total est		
	69,7	80,2	
macroformes de la surface	6	17	
orientation	7	52	O-S-E et NO-N-NE
	5	18	O-SO, SE-E et NO-N-NE
altitude	10	22	
végétation	10	14	
Facteurs de relief	Valeur de la variation de la densité : $\Delta \gamma = \gamma_{max} - \gamma_{min}$ kg/l si la densité moyenne dans le bassin versant total est		Remarques
	0,170	0,230	
macroformes de la surface	0,02	0,02	
orientation	0,045	0,035	O-S-E et NO-N-NE
	0,005	0,02	O-SO, SE-E et NO-N-NE
altitude	0,01	0	
végétation	0,01	0,02	

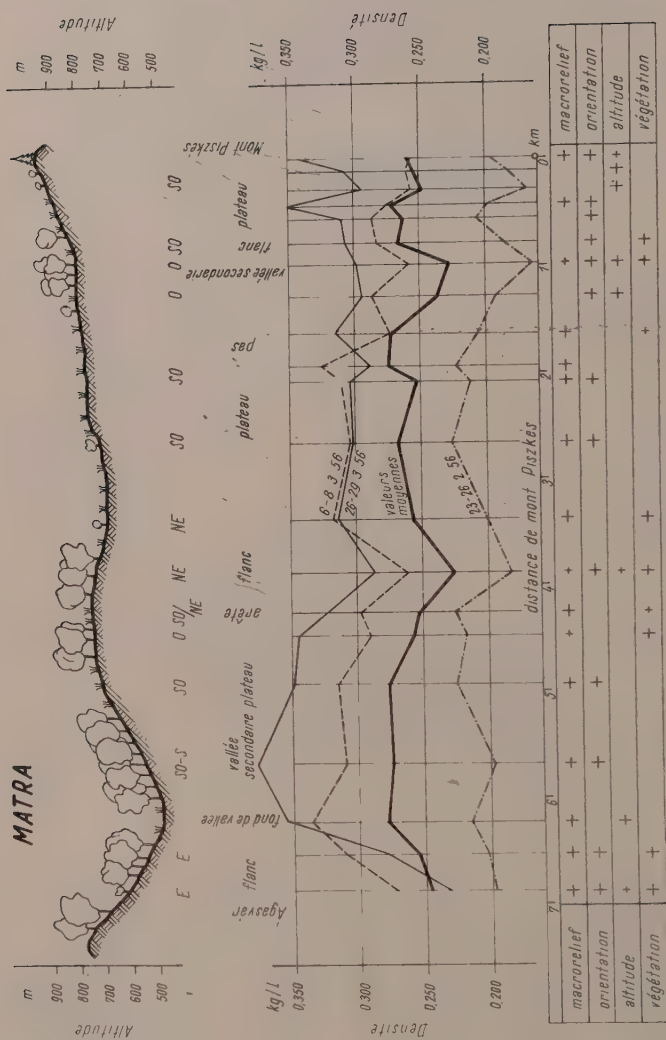


Fig. 7 — L'influence de l'ensemble des facteurs de relief sur la densité le long du cheminement. Monts Mátra.

de rares taches dans les Mtára). *L'orientation du bassin versant ne représente qu'une influence relativement faible* si nous comparons l'ensemble des superficies situées vers les O-S-E et correspondant au secteur de 225°, avec les superficies orientées vers les NO-N-NE et correspondant à un secteur de 135°. Par contre, si l'on ne comparait que les superficies situées vers le Nord et celles exposées au Sud, l'examen aboutirait à l'influence la plus importante équivalant ou dépassant même celle de la situation par rapport à l'altitude.

Dans le Tableau 5, nous reproduisons les résultats des observations effectuées dans les *Monts Bükk* pour le cas où la densité moyenne est $\gamma = 0,17$ et 0,23. Ces données ne caractérisent que les conditions d'accumulation et de fusion de quelques parties abruptes des Massifs centraux hongrois. *L'influence de l'orientation* et, en particulier, celle des flancs exposés au Sud sont significatives. Ce qui est dans la nature des choses puisque dans le bassin versant du Garadna, le nombre des flancs situés vers le Sud, dépasse les 20% de la superficie. *L'influence de l'altitude* paraît moins important dans la variation de la densité, mais on peut la démontrer pourtant avec certitude dans la variation de la teneur en eau. *Les macroformes* du relief exercent une action à peu près analogue dans la vallée de la Garadna et dans les *Monts Mátra*.

Pour permettre d'apprécier *l'influence conjuguée de l'ensemble des facteurs de relief*, dans la Figure 7, nous représentons les changements de la densité de neige pendant trois périodes d'observation différentes. Ces observations avaient été effectuées dans les *champs d'expériences des Monts Mátra* sur l'un des cheminement, choisis de manière à couvrir toute la région étudiée. Le facteur de relief et le facteur de végétation intervenant d'une manière décisive pour déterminer la valeur de la densité, ont été reproduits séparément aussi. En étudiant la Figure 7, on se rendra facilement compte de l'influence de l'ensemble des facteurs. Ici encore, on se contentera de souligner seulement quelques faits à titre d'exemple. Sur le plateau de Piskés, l'influence du relief (haute plaine) et celle de l'orientation (SO) tempèrent l'action exercée par l'altitude. Dans le champ d'observation orienté vers le NE et situé à 3,9 km du plateau de Piskés, la valeur très basse de la densité résulte de l'action concertée de trois facteurs : relief, orientation, végétation. Alors que dans la partie inférieure de la vallée de Csörgő, l'influence du relief (fond de vallée) et celle de l'altitude (la vallée en question représente la plus basse altitude parmi les points d'observation reproduits dans la Figure 7) donnent une densité élevée.

Nous avons fait la tentative d'appliquer *le calcul des probabilités* à l'examen de l'influence séparée de chacun des facteurs de relief (Tableau 6). L'expérience n'a pas donné un résultat aussi concluant que nous avait fourni ce calcul appliqué préalablement à l'ensemble des données de mesure [9]. Cela n'est pas pour surprendre. Quand on veut calculer la probabilité d'un détail séparé de l'ensemble, on ne peut disposer d'un nombre de cas suffisamment élevé et assurer la répartition des données exigée par la théorie des chances. Pourtant, le calcul des probabilités nous a permis, dans ce cas aussi, de fixer plusieurs faits dont nous allons citer quelques exemples. Les facteurs d'asymétrie tirés par déduction des données relatives aux plateaux, représentent toujours la valeur la plus élevée ce qui ressort même de la situation variée des plateaux. Cette règle est valable pour les superficies orientées vers les O-D-E également. La valeur des facteurs d'asymétrie ne descend au-dessous de celle des superficies exposées aux NO-N-NE que dans la dernière période de la fusion. Ce qui souligne encore l'aspect varié des terrains orientés vers le Sud. Pour les superficies dont l'altitude reste au-dessous de 650 m, le facteur d'asymétrie atteint déjà la valeur négative, la fusion étant plus rapide à ces endroits. Les facteurs de variation des hautes plaines, des superficies orientées vers les O-S-E et des terrains en dessous de 650 m sont, à l'exception d'un seul cas, plus élevés que ceux des autres superficies. Ce qui rappelle encore le caractère accidenté de ces terrains.

* * *

TABLEAU 6

Calcul des probabilités de l'influence des facteurs de relief

Facteurs de relief	Périodes de mesures								
	23-26. II				6-8. III				
	Valeurs des facteurs de probabilité								
	M	C _v	C _s	M	C _v	C _s	M	C _v	C _s
vallée secondaire	0,217	0,099	-0,43	0,318	0,090	-0,05	0,345	0,091	+0,14
flanc	0,204	0,087	+0,20	0,291	0,092	+0,49	0,318	0,121	-0,12
plateau	0,220	0,144	+1,56	0,309	0,140	+0,58	0,333	0,097	+0,65
O-S-E	0,221	0,135	+1,34	0,306	0,112	+1,37	0,326	0,115	-0,28
NO-N-NE	0,212	0,097	+0,61	0,299	0,093	+0,64	0,325	0,101	+0,40
au-dessous de 650 m									
au-dessus de 650 m							0,336	0,117	-0,33
Remarques	M	valeur moyenne					0,320	0,091	+0,64
	C	facteur de variation							

Les résultats exposés ci-dessus semblent présenter un double intérêt : d'une part, ils permettront d'établir à des endroits judicieusement choisis dans les montagnes hongroises, des stations de mesure pour l'observation suivie de la neige; d'autre part, ils fournissent un instrument de recherche commode pour évaluer, sur la base des données relevées par ces stations, le volume d'eau emmagasiné dans les bassins versants de ces montagnes.

Nous remarquerons, pour terminer, que dans notre étude nous n'avons pas examiné tous les problèmes en rapport avec le relief. Ainsi, par exemple, en développant les résultats obtenus dans les montagnes, nous avons passé sous silence l'influence directe de l'angle de la pente des versants. Nous n'avons pas abordé, non plus, l'étude de l'influence du relief dans la plaine et la région des collines. L'examen de ces problèmes aurait exigé un grand nombre de mesures dont nous n'avons pas disposé ou aurait débordé le cadre de notre étude. Malgré ce fait, nous espérons d'avoir contribué à éclaircir l'influence des facteurs de relief.

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ON THE CALORIMETRIC DETERMINATION OF SNOW QUALITY

U. RADOK, S.K. STEPHENS and K.L. SUTHERLAND

SUMMARY

The classical method of determining snow quality works satisfactorily only when great care is taken to melt the snow electrically at ambient temperatures. As a superior alternative, especially for field use, a reversed procedure is suggested in which the snow sample is frozen rather than melted. It is shown that much greater accuracy can be obtained in this way for the same precision of measurement, and a technique suitable for field use is described.

1. INTRODUCTION

Among the various methods suggested for the measurement of snow quality, since Bernard and Wilson (1941) first pointed out its significance, their calorimetric approach retains a special place. This is because it alone shows directly the effect of the free water in the snow on the melting process. However, the use of hot water as in the original form of the calorimetric method leads to errors which tend to produce a spurious increase in free moisture content with snow sample size (Radok 1956), an effect already evident in the data quoted by Bernard and Wilson (1941). As a remedy K. Croce (cf. Halliday 1950) suggested the electric melting of the snow at ambient temperature. One of us (Stephens) has obtained satisfactory results with such an arrangement. The measurements are, however, time-consuming and require more electric energy than is readily available in the field. As an alternative another of the present authors (Sutherland) determined the thermal quality of the snow directly through its behaviour in *freezing* rather than melting. It will be shown that this has both theoretical and practical advantages.

2. PROCEDURES

In the form developed by Stephens the electric melting calorimeter consisted of a wide-mouthed Dewar flask of 1 litre capacity with a large heating coil operating on 230 volts from batteries and a rotary converter. The current passed through a watt-hour meter calibrated to within 0.02 watt hour (17 calories). The high voltage, made possible by the facilities of the Spencer's Creek research station of the Snowy Mountains Hydroelectric Authority, reduced the time required for the melting of 150-200 gm. of snow in 400 gms of water at around 4°C to about 10 minutes. Against this a portable version of the electric calorimeter described by Hansen and Jellinek (1957) operates on 6V and requires more than 1 hour to melt 10 gms of ice. This illustrates the main drawback of the electric melting method, especially under field conditions. There is little doubt, however, that it is invaluable as a standard of reference when used under semilaboratory conditions as in the present case.

The freezing calorimeter technique of Sutherland used the same type of Dewar flask which was filled with about 250 cm³ of iso-octane (2,2,4-trimethylpentane) or petrol, precooled to between -15°C and -25°C by a freezing mixture prepared from CaCl₂·6H₂O and snow in the proportion of 1.2 to 1 by volume. The size of the snow sample inserted into the iso-octane was chosen so as to bring the final temperature of the mixture to just below 0°C. It proved desirable to inject the snow in batches of 50 cm³ or less since a larger sample of snow (especially when wet

tended to freeze in a single lump, trapping the free water and insulating it against the colder iso-octane. The low heat capacities of both snow and iso-octane (given as functions of temperature in table 1) and the differences between the calorimeter and ambient temperatures made it necessary to observe cooling curves in this case. All quantities were measured by weighing with the field balance of the type used by the Canadian Snow Survey (Pearce and Gold 1951)(*) to simulate field working conditions.

TABLE 1

Specific heats of ice and iso-octane (in 15°C cal.)

Temperature	Ice cal/g/°C	Iso-octane cal/g/°C
-20°	.464	.450
-15°	.472	.455
-10°	.479	.460
- 5°	.486	.465
0°	.493	.470

3. THEORY

In discussing the heat balance for electric *melting* calorimetry we shall assume that the final temperature t_0 is identical with the initial temperature of the water in the calorimeter; the extension to slightly differing initial and final temperatures is of course straightforward. Then the heat balance relation is

$$L(S - F) + t_0S = H \tag{1}$$

where S and F are the weights of snow and free water, both initially at 0°C, t_0 is the constant calorimeter temperature, L is the latent heat of melting, and H is the heat added electrically. The snow quality Q_m follows from (1) as

$$Q_m = 1 - F/S = H/SL - t_0/L \tag{2}$$

In *freezing* calorimetry we start with W_i grams of iso-octane of specific heat c_i (cf. table 1) in the calorimeter of "iso-octane equivalent" E at the initial temperature t_1 ; for simplicity $W_i + E$ will be denoted by W'_i . Giving S , F , and L their previous meaning, and with c_s denoting the specific heat of ice (cf. table 1) and t_2 the final temperature in the calorimeter, the heat balance is

$$W'_i c_i (t_2 - t_1) + t_2 c_s S = LF \tag{3}$$

so that the snow quality as determined by freezing calorimetry is

$$Q_f = 1 - W'_i c_i (t_2 - t_1)/LS - c_s t_2/L \tag{4}$$

(*) The authors are indebted to Dr. L. W. Gold for the drawings of this balance.

The last term in both (2) and (4) represents merely a small correction. Apart from this (2) contains a single large term of order unity which in (4) is obtained through a small difference from unity. In physical terms, the melting procedure compares the heat needed in absence of free water with that actually needed for the melting of the snow, two large and nearly equal quantities. Against this in the freezing procedure the much smaller amount of heat released in the freezing of the free water is measured by itself and compared with that required to melt the snow in absence of free water. This implies a higher intrinsic accuracy for freezing calorimetry.

Detailed error considerations for the two experimental procedures described in section 2 indicated that the electric calorimeter should give results correct to within $\pm 2\%$ of free water, while the freezing method operating with cruder field equipment should be accurate to within $\pm 1\%$ of free water.

TABLE 2

Comparison of free water contents obtained with electrical and freezing calorimeters from snow samples taken near Spencer's Creek station, Mt. Kosciusko.

Determination	Grain size	% Free Water		Remarks
		Electical	Freezing	
1	2-3 mm	3.2	4.5	Large grains 20'' below snow surface
2		2.4	5.6	
3		10.1	9.6	
4		9.6	9.2	
5		—	10.6	
6	2-3 mm	5.2	3.4	moist granular surface snow early morning
7		5.1	3.3	
8		4.8	3.6	
9	1 mm	25.0	22.9	fine grain surface snow, 1 p.m.
10		24.3	24.2	
11	$\frac{1}{2}$ -1 mm	23.7	23.0	fine grain surface snow, 2 p.m.
12		22.9	27.8(*)	
13	2-3 mm	3.5	3.1	coarse grain surface snow 4 p.m.
14		27.2	20.4(*)	very wet surface snow 4 p.m.
15		26.1	26.4	
16	$1\frac{1}{2}$ mm	16.6	12.4	surface snow in snow drift.
17		13.1	11.8	
18		17.6	14.9	

(*) sampling uncertain. Brackets indicate replicate measurements.

4. RESULTS

A few figures obtained during the original comparisons of the two methods at Spencer's Creek near Mt. Kosciusko in 1958 are given in table 2. Strict agreement within the estimated limits is ruled out by the large local variability of snow quality, especially in the moist surface layer; thus in the context of natural snow the figures in table 2 demonstrate that the two methods give equivalent results. The simple freezing technique of Sutherland can therefore be regarded as a direct and accurate method of measuring snow quality in the field.

5. ACKNOWLEDGEMENT

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PRECISION DES MESURES D'ABLATION

A. BAUER (EGIG)

RÉSUMÉ

Un nouveau type de balise d'ablation a été expérimenté avec succès au Groenland pendant la Campagne d'été 1959 de l'EGIG. Les mesures de l'ablation nette montrent que l'erreur maximum avec une seule balise peut être de l'ordre de 26 % rendant illusoire tout calcul du bilan de masse. La précision peut être augmentée en augmentant le nombre des balises de mesure, chacune étant elle-même contrôlée par cinq ou dix balises supplémentaires.

SUMMARY

A new type of ablation stake was tested successfully in Greenland during the Summer Campaign 1959 of the EGIG. The measure of net ablation shows that the maximum error may be about 26 % : that shows quite dilusive every estimate of mass balance. Precision may become better by increasing the number of measure stakes, each of them being controlled by 5 or 10 supplementary stakes.

1. INTRODUCTION

L'E.G.I.G. (Expédition Glaciologique Internationale au Groenland) est une expédition commune aux pays suivants : Allemagne, Autriche, Danemark, France et Suisse. Elle fut créée en 1956 (Bul. A.I.H.S., n° 2, 1956). Elle fut autorisée par le Gouvernement du Danemark et reçut le patronage de l'Association Internationale d'Hydrologie Scientifique et sa Commission des Neiges et des Glaces. Son organisation et son programme ont été publiés (Bul. A.I.H.S., n° 6, 1957).

L'E.G.I.G. est placée sous la responsabilité d'un Comité de Direction dont le Bureau actuel est le suivant :

Président : Dr. B. Fiistrup (Danemark)

V. Présidents : Prof. Finsterwalder (Allemagne) et Prof. Kobold (Suisse)

Représentant du Gouvernement Danois : Dr. H. Larsen

Chef d'expédition : P.E. Victor (France)

Secrétaire Général : Prof. A. Bauer (France)

Toute la réalisation technique de l'EGIG a été confiée aux Expéditions Polaires Françaises dont le Directeur, P.E. Victor est en même temps le chef de l'EGIG. Un très important support aérien a été réalisé par l'Armée de l'Air Française.

Après deux années de préparation et de reconnaissance (1957-1958), le programme de recherches glaciologiques dans la partie centrale du Groenland a été réalisé pendant la campagne d'été de 1959. L'EGIG se terminera en 1960 par une campagne réduite ayant surtout pour mission de ramener le matériel et d'évacuer les six hivernants de la Station JARL-JOSET.

Les résultats scientifiques de l'EGIG seront publiés dans la revue danoise MEDDELELSER om GR ØNLAND.

2. E.G.I.G. : GROUPE DE GLACIOLOGIE CÔTIÈRE

Le Groupe de Glaciologie Côtière de l'EGIG a été installé le 10 mai 1959 au Camp IV EGIG par hélicoptère. Tout le matériel (12 tonnes) a été reçu par air drop. Le Groupe a quitté la Station le 11 août 1959.



Fig. 1

Le Groupe se composait de trois personnes : A. Bauer, chef de groupe, W. Ambach (climatologie et bilan de radiation) et O. Schimpp (glaciologie).

Le groupe avait pour mission :

- de déterminer le mouvement horizontal et vertical de la surface de la zone d'ablation sur un profil de 80 km par triangulation et nivellement géodésique;
- d'étudier le mouvement du bord de l'indlandsis et du glacier Eqip Sermia;
- d'étudier l'ablation et l'écoulement de l'eau de fonte;
- d'étudier la glace surimposée et la glaciologie superficielle;
- d'effectuer toutes les mesures nécessaires à la climatologie et au bilan de radiation.

La climatologie et les mesures et enregistrements pour déterminer le bilan de radiation ont été effectués par W. Ambach à la station. Tous les autres travaux ont été effectués par A. Bauer et O. Schimpp. A part un léger support par hélicoptère en début et en fin de campagne, tous les déplacements ont été effectués à pied.

3. LES MESURES D'ABLATION: POSITION DU PROBLÈME

L'ablation de la glace est mesurée par la diminution de la glace par fusion par rapport à un repère fixe constitué par une balise profondément enfoncée dans la glace.

Je me bornerai dans le présent travail à la seule mesure de l'ablation nette. Je ne parlerai pas de la glace surimposée, problème qui ne se pose pas pour l'exemple choisi.

La surface d'un glacier en général, et celle de la zone d'ablation au Groenland en particulier, est loin d'être uniforme. Elle varie de nature de dcm en dcm : vieille glace pleine de bulles d'air, vieille glace souillée par la cyoconite, bandes bleues de différentes espèces, bédrières, etc. Cette variété de la surface conduit à une ablation différente de place en place. Cette ablation différentielle se manifeste par la formation de bosses et de creux pouvant avoir une dénivellée de plus d'un mètre.

Le problème est donc de savoir avec quelle précision l'ablation peut être mesurée en ne tenant pour le moment uniquement compte que de la nature de la surface. Ce problème a déjà été mentionné, mais sans recevoir une solution définitive (Liestøl, 1954).

4. EXEMPLE DE MESURES D'ABLATION

L'exemple choisi est constitué par les mesures d'ablation C.

Les balises C ont été implantées au voisinage de notre station (Camp IV EGIG, 69°40' N, 49°37' W, 1 000 m). Il s'agit du sommet d'une colline, crevassée légèrement et parsemée de bosses de glace blanche. Ces bosses provenaient de l'ablation différentielle de la glace blanche entre des bandes de glace bleue inclinée à 45° environ dans la masse de la glace. Le sommet de cette colline était libre de neige avant la fonte estivale et il ne s'y est pas formée de glace surimposée. La période d'observation s'étend du 5 mai au 8 août 1959. L'ablation mesurée est donc uniquement de l'ablation nette.

4.1. *Types de balises utilisées*

4.1.1. *Balise C6 type Kasser*

La balise principale C6 était du type employé par P. Kasser en Suisse. Elle était constituée par des éléments de deux mètres en bois de sapin de 3 cm de diamètre. Les éléments étaient reliés les uns aux autres par des chainettes. Un manchon en caoutchouc les rendait solidaires les uns des autres. Deux lames de ressort en acier fixées au bas de chaque élément empêchaient la balise de remonter dans le trou de forage et la fixaient dans la glace. Comme la glace est à une température négative en-dessous de -1 m, la balise était en plus prise dans la glace très peu de temps après la pose. La balise C6 était constituée par 5 éléments de 2 m, soit une longueur totale de 10 m.

Par suite du rayonnement, il se formait autour de la balise un trou d'environ 15 cm de diamètre et profond d'environ 60 cm. Le bord de ce trou était très irrégulier, ce qui rendait imprécis la mesure de la partie de la balise au-dessus de la surface de la glace, au moins pour des mesures diurnes de l'ablation nette.

4.1.2. *Balises fil — C1 à C5*

Cette balise C6 devait être contrôlée par un ensemble de balises supplémentaires. Pour économiser le poids à transporter, mais aussi pour augmenter la précision des lectures journalières, nous avons utilisé un nouveau type de balise. Comme la glace était à une température négative, et comme l'ablation estivale était de l'ordre de plus



CAMP IV. - E.G.I.G

Balises C

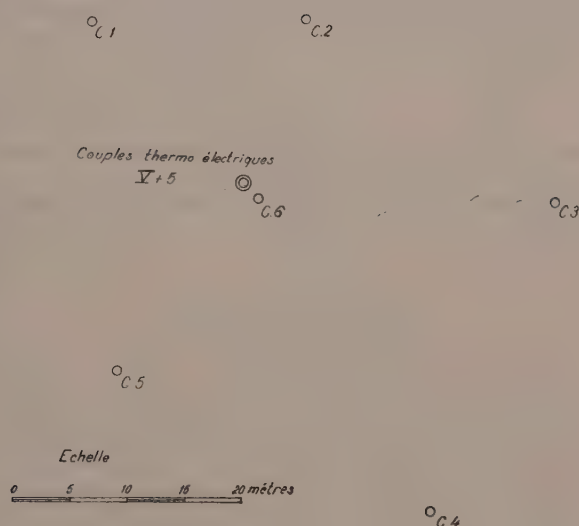


Fig. 2

d'un mètre, nous n'avons pas pu employer la méthode de H. Hoinkes : simple trou foré dans la glace, le fond du trou constituant le repère fixe (Hoinkes, 1954).

La balise était constituée par un câble en acier inox tressé de 30/100 e de diamètre et d'une résistance de 9 kg. L'extrémité inférieure de la balise d'une longueur d'environ 5 m était lestée par un boulon de 30 gr environ. A partir du bas, aux horizons 100, 200, 250, 300, 350 et 400 cm, un petit bouchon de pêche au brochet était fixé sur le câble. Le haut du câble portait un noeud et un petit fanion.

Le câble était introduit dans le trou de forage rempli d'eau et profond d'environ 4,50 m. Le boulon se posait au fond et les bouchons maintenaient le câble absolument vertical et sans aucune boucle. Au bout de quelques minutes, le boulon était pris dans la glace par suite du regel de l'eau du trou. Toute la balise était prise dans la glace au bout d'une demie-heure. Nous avions ainsi un câble vertical fixé dans la glace. Le point repère était constitué par le noeud à l'extrémité supérieure, facile à trouver grâce au fanion. On mesurait la hauteur de ce noeud au-dessus de la surface de la glace. Cette mesure était très précise, car le câble était pris dans la glace à la surface et ne de la donnait pas un grand trou par rayonnement comme pour la balise en bois.

Ce type de balise, déduite de la technique de la pêche, s'est révélé très pratique. Il était aisé de transporter 600 m de balise de ce type dans sa poche.

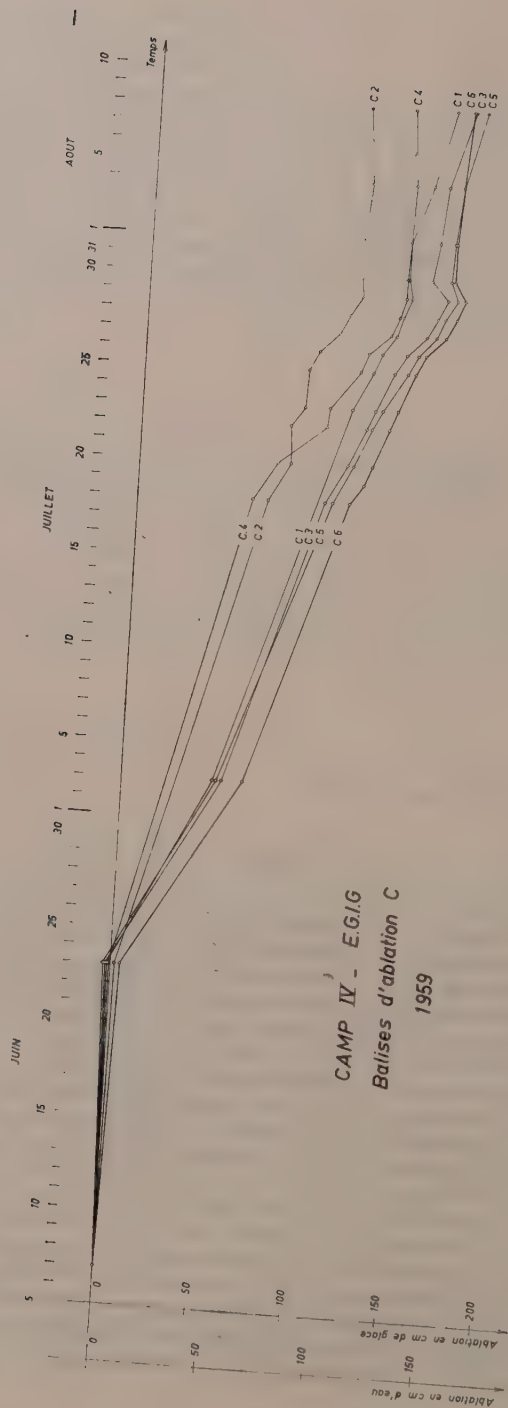


Fig. 3

4.2. Implantation des balises

Les trous de forage ont été réalisés à l'aide d'une sonde thermique à eau chaude prêtée par P. Kasser (Kasser, 1951). Cette sonde, d'un poids total d'environ 80 kg avec accessoires, permettait de forer des trous de 4 cm de diamètre jusqu'à 30 m de profondeur avec une vitesse de l'ordre de 12 à 16 m à l'heure. Le liquide circulant en circuit fermé était un mélange d'eau et d'éthylène glycool à 50% ne pouvant geler aux températures les plus basses rencontrées dans la glace (-10°C). Au cours Campagne d'été de l'EGIG, nous avons foré environ 400 m au total.

Les 5 balises fil C1 à C5 ont été régulièrement disposées autour de la balise en bois C6 le long d'une circonférence d'environ 25 m de rayon. Les balises C1 à C5 se trouvaient ainsi dans des conditions superficielles très différentes, sommet d'une bosse, creux, creux avec bédrière, etc. Les mesures étaient en général effectuées vers 21 h.

4.3. Résultats des observations

Les résultats des observations se trouvent dans le tableau ci-contre et dans le graphique.

Le deuxième tableau donne la valeur moyenne de l'ablation et les diverses précisions.

*Balises fil C1 à C5
Balise type Kasser C6*

Valeur de l'ablation cumulée en cm de glace

Date	C1	C2	C3	C4	C5	C6
050659						00
070659	00	00	00	00	00	00
230659	+ 02	02	+ 03	+ 01	+ 04	05
030759	48		53		50	64
180759		69	99	61	103	112
190759						119
200759		80	110	73	113	123
220759		79	119	98	122	131
230759	111	86	123	99	127	135
250759	121	87	132	114	139	143
260759	125	92	138	118	144	148
270759	132	102	148	129	153	158
280759	135	107	153	133	157	163
290759	139	113	158	136	163	167
300759	137	112	149	136	159	161
010859	137	112	152	137	160	161
040859	147	115	155	138	163	163
080859	157	112	167	135	173	166

Valeur moyenne de l'ablation et précision

Balise	ablation glace cm	écart e	e ²
C1	157	- 5	25
C2	112	+ 40	1600
C3	167	- 15	225
C4	135	+ 17	289
C5	173	- 21	441
C6	166	- 14	196
	910		2776

Valeur moyenne de l'ablation 152 cm
 Erreur moyenne d'une seule détermination $\pm 23 \text{ cm} = \pm 15\%$
 Erreur moyenne quadratique $\pm 10 \text{ cm} = \pm 6\%$
 Erreur maximum d'une seule détermination $\pm 40 \text{ cm} = \pm 26\%$

5. CONCLUSIONS

5.1. Mesure de l'ablation avec une seule balise

Mesurer l'ablation avec une seule balise pour une zone relativement importante est une opération illusoire. L'erreur maximum peut atteindre 26%.

Si nous reprenons les valeurs publiées pour établir le bilan de masse de l'indlandsis groenlandais (Bauer, 1954), nous trouvons une incertitude maximum de 80 km³ pour une ablation nette totale de 315 km³. Il est donc impossible de parler de bilan de masse de l'indlandsis. Cette incertitude doit dans ce cas être aussi grande lorsque la même méthode est utilisée pour un glacier quelconque.

Pour améliorer la précision des mesures d'ablation, il faut donc implanter beaucoup de balises dans des zones différentes, chaque balise étant en plus contrôlée par 5 ou dix balises supplémentaires : c'est ce que nous avons réalisé pendant la Campagne d'été 1959 de l'EGIG. Le compte-rendu de tous ces travaux dépasse le cadre de cet article et sera publié ultérieurement.

5.2. Mesure avec plusieurs balises

La mesure de l'ablation nette avec une balise contrôlée par cinq balises supplémentaires permet une précision de l'ordre de 6%. Cette précision est nettement meilleure que la précédente. Elle reste néanmoins insuffisante.

Comme conclusion générale, pour obtenir une bonne précision de l'ablation il faut donc distribuer les balises en réseau serré sur le glacier à étudier, chaque balise étant contrôlée par un nombre important de balises supplémentaires. De plus, comme l'ablation annuelle est variable, il faut implanter des balises permettant les observations sur une période d'au moins dix années pour obtenir une valeur de l'ablation moyenne annuelle valable et représentative.

Les courbes du graphique ne peuvent continuer à diverger. Les bosses de glace doivent à un moment donné disparaître et peut être faire place à des creux, mais cette évolution prend plus d'une saison d'ablation dans la région étudiée, aussi bien par l'ablation elle-même que par le mouvement de la masse de la glace.

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APPLICATION OF THE HYDROCHEMICAL METHOD FOR INVESTIGATING THE RIVER RUN-OFF FED BY SNOW GLACIERS

V.L. BLINOVA (U.S.S.R.)

SUMMARY

I. When a thaw is in a top the daily flow of snow -glacial feeding- river may be express by equation :

$$Q = q_{mar} + q_{nv} \pm q_{2p} + q_{goms}. \quad (1)$$

where

Q total daily flow of a river

q_{mar} thawed waters flowing in the river-bed from a glacier surface.

q_{nv} thawed waters soaked on the glacier-bed from surface and waters formed as a result of the underglacial ablation.

q_{goms} flow of liquid precipitations which is small quantity for the most of Alpine waters.

q_{2p} subsoil waters.

II. The subsoil feeding of mountain rivers has a variable sign for twenty-four-hours : whenever a high-water flows the hydrostatic pressure in the river-bed increase and water-bearing layers (or the whole of the loose-ground forming the river-bed) saturate with a slightly mineralized thawed water. When maximum daily flow expenditure is practically equal to

$$Q = q_{mar} + q_{nv} \quad (2)$$

As a high-water decreases a hydrostatic pressure falls down and absorbed considerably mineralized waters enter in action. Subsol feeding reaches its maximum when a daily flow has a minimum.

III. As a rule, snow-glacial feeding rivers have the steady (night and morning) minimum flows.

It's especially match for the rivers situated near glaciers. These flows fall on the time when the flowing from surface does not occur. During such periode the flow is equal practically.

$$Q = q_{nv} + q_{2p} \quad (3)$$

The constancy of the minimum flows prove that the sources of the rivers feeding are regular.

IV. A water mineralization is different for each of the three fundamental flow components. The degree of river waters mineralization depends on the predominance of one of the components in daily flow and is inversely proportional to the daily course of the flow. At any rate the mineralization of the thawed and underglacial water during short periods is constant and may be easily determined.

V. The curve of functions $A = f(Q)$ for about twenty-four-hours permit to mark out the periods when some components prevail in flow.

The curves $A = f(Q)$ (where A is the total water alkalinity) have two distinctly expressed branches corresponding to the periods of raising and abatement.

VI. Knowing the mineralization of thawed waters on the grounds of the curves $A = f(Q)$ one can determine the value of the underglacial flow which is practically constant for twenty-four-hours. In that way with the help of the equation (3) one can also determine the value of the subsol flow.

VII. The method described above provides the possibility of the quantitative examination of the regulation role of the underglacial capacities for the flow of snow-glacial feeding rivers.

RÉSUMÉ

1. L'écoulement quotidien des fleuves alimentés par les nieges et glaciers au plus fort de la fonte peut être exprimé par l'équation :

$$Q = q_{fonte} + q_{in} + q_g + q_s \quad (1)$$

où

Q — écoulement total quotidien du fleuve,

q_{fonte} — eaux de fusion, ruisselant dans le lit directement de la surface du glacier,

q_{in} — eaux de fusion, s'infiltrant de la surface du glacier dans son lit, et eaux

formées par suite de l'ablation de la partie sous-glaciaire,

q_g — écoulement dû aux précipitations liquides, valeur négligeable pour la majorité des régions de haute montagne,

q_s — eaux souterraines.

2. L'alimentation des torrents en eaux souterraines est précédé d'un signe alternatif au cours des 24 heures de la journée : lors des grandes crues la pression hydrostatique dans le lit s'accroît et il se produit une saturation des lits aquifères (ou de toute l'épaisseur poreuse composant le lit) en eau de fusion faiblement minéralisée.

Le débit quotidien de l'écoulement est pratiquement égal à :

$$Q = q_{\text{fonte}} + q_s \quad (2)$$

3. A mesure que les grandes crues s'abaissent, la pression hydrostatique tombe et les eaux absorbées, déjà enrichies de substances minérales, entrent en action. L'alimentation souterraine atteint son maximum avec un débit minimum.

4. Pour les fleuves alimentés par les neiges et glaciers des débits stables minimum (nocturnes et du matin) sont caractéristiques. Ces débits ont lieu lorsque cesse tout écoulement de la surface du glacier. Pratiquement l'écoulement durant ces périodes s'exprime par :

$$Q = q_{\text{in}} + q_s \quad (3)$$

La stabilité des débits minimum témoigne de la régularité de ces sources d'alimentation.

5. La minéralisation des eaux diffère avec chacune des trois sources d'alimentation principales. La modification quotidienne de la minéralisation des eaux du fleuve est fonction de la prédominance de l'une de ces sources dans l'écoulement et inverse au débit quotidien.

La minéralisation des eaux de fonte et eaux sous-glaciaires est en tout cas constante pour une courte durée et peut être facilement déterminée.

6. fonction du type $H=f(Q)$ donne des courbes qui permettent de dégager nettement les périodes de prédominance de chacun des composants du débit. Les courbes $A=f(Q)$ (où A — alcalinité de l'eau) ont deux branches clairement exprimées qui correspondent aux périodes d'augmentation et de diminution de l'écoulement.

7. On peut déterminer, à l'aide des courbes $A=f(Q)$, la minéralisation des eaux de fusion étant connue la valeur du débit des eaux de fusion, valeur demeurant pratiquement constante durant les 24 heures de la journée. De même, en partant de l'équation (3) on peut déterminer la valeur du débit des eaux souterraines q_s .

8. La méthode décrite permet une détermination quantitative du rôle régulateur des capacités constituées par les parties inférieure et intérieure des glaciers dans le débit des fleuves alimentés par les neiges et glaciers.

Experiments on the investigation of river run-off fed by snow glaciers were carried on during the summer of 1959 on the Southern slope of the Elbrus. The object for investigation was the Garibashi River flowing from under the glacier of the same name. The territory, on which observation was carried on, is situated above any of the tributaries, therefore the river is fed only at the expense of the Garibashi glacier and by the snow fields pertaining to it.

The daily river run-off fed by snow glaciers, during the period of snow thawing, may be expressed by the following equations:

$$Q \text{ (per day)} = q_1 \text{ (melt)} + q_2 \text{ (subglacial ablation)} \pm q_3 \text{ (ground waters)} + q_4 \text{ (rain waters)}$$

where q_1 — represents melt waters flowing from the glacier surface directly into the riverbed.

q_2 — waters finding access to the glacier surface, inside it and onto its bed (their arrival into the river bed being therefore retarded), as well as waters formed as a result of subglacial ablation.

q_3 — ground and intermorainic waters.

q_4 — the runoff of liquid precipitation, the volume of which is negligibly small

in the vicinity of the glacier and is therefore herein disregarded. The role of the enumerated components during a 24-hour period is not similar. During active thawing the volume of q_1 — constitutes the predominant amount of waters in the river bed. During night time the volume of q_1 decrease down to zero.

The run-off of subglacial waters (q_2) is kept constant by the recesses found on the body and bed of the glacier (cracks, cavities, etc.) and remains unchanged within a brief period of time.

Ground water component (q_3) is changeable. When floods begin, the hydrostatic pressure on the river bed increases and saturation of the water-bearing strata or of the entire loose strata forming the river bed with melt water takes place. As the water level decreases, the pressure on the river bed also decreases and the yield of "stored" ground waters to the river bed begins. The ground water runoff reaches its maximum capacity, when melt is at the lowest, i.e. when there is no run off from the surface of the glacier (which is usually observed during the second half of the night or in early morning). It should be noted, that the minimal river level changes have a comparatively smooth character, forming, so to say, "basic" levels, which carry the daily maximum peak waters. When the average temperature in the region of alimentation remains low during several nights (i.e., when the run-off is made up of ground and subglacial waters), the minimal level remains at a constant mark, regardless of day-time fluctuations.

Thus, the daily discharge of waters in the river during melting may be expressed by the equation:

$$Q \text{ daily} = q \text{ melt} + q \text{ subglacial ablation}$$

while the night-time minimum discharge equals:

$$Q \text{ nightly} = q \text{ subgl. ablation} + q \text{ ground w.}$$

The changing content of dissolved substances in the water may serve as an indirect indication of the particular prevailing component in the runoff, inasmuch as the mineralization of each of these components differs from one another.

The chemical composition of the ice supplying the main daily volume of runoff depends upon the composition of the rocks surrounding the glacier and on the atmospheric precipitation; it is sufficiently constant and easily defined. The waters of the subglacial runoff become more strongly mineralized during their contact with the glacier bed, but their mineralization is also not affected by sharp fluctuations with respect to time. Ground waters constitute the most labile and difficult to define part of the chemical river runoff. The poorly mineralized melting waters injected by the flood into the water-bearing strata return to the river bed sufficiently enriched in salt content.

The varying content of salts in the water maintains a course which is contrary to that of the daily runoff (Fig. 1). The definition of the salt content and its variation in the course of 24 hours may be made on the basis of one of the elements, giving the best reflection of the total daily change. On the Garibashi River the definition was based on the content of silicon and the general alkalinity. The amount of silicon content in the water proved very low (somewhere around 0.003 mg/l), and in the future only the general alkalinity was determined by direct titration with muriatic acid. The general alkalinity in conditions of the Elbrus lava streams proved to be quite considerable (up to 0.65 mg/l.p.m. and gave a good picture of all the runoff fluctuations (Fig. 1). Tests were made every two hours during day and night, the nightly tests (those taken between 8 P.M. and 6 A.M.) were analysed no later than four hours after sample taking, while those taken during daytime hours (from 8 A.M. to 6 P.M.) were analysed immediately.

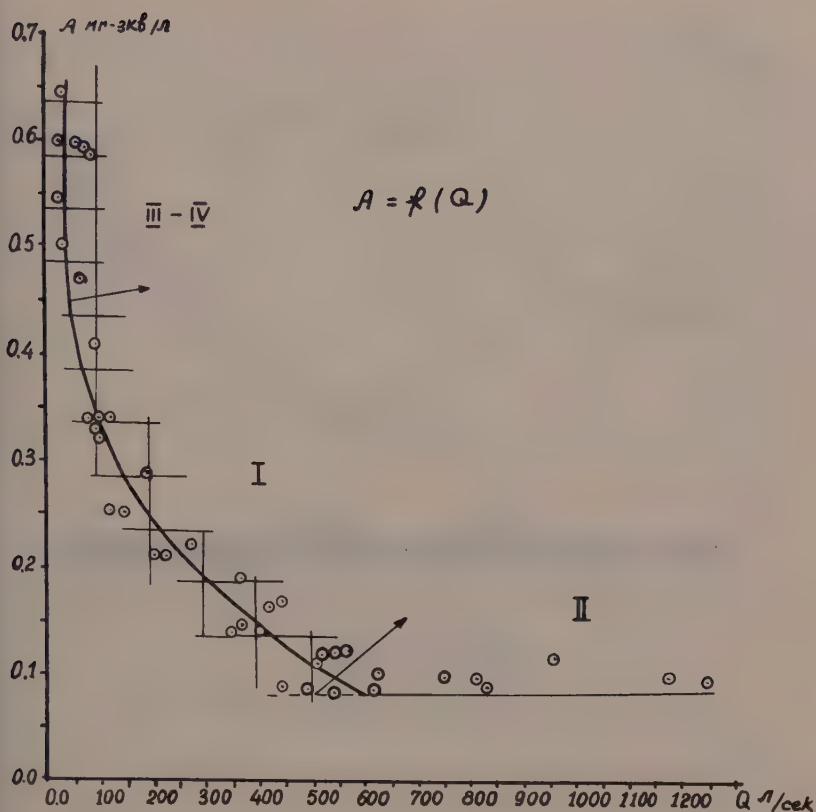


Fig. 1 — Dependence of common alkalinity on the expense of water

Between the discharge of water its general alkalinity there exists a clearly defined inverse proportion. The curves $A = f(Q)$ have two clearly expressed branches of rise and fall (Fig. 3). It is characteristic that in all our series of alkalinity tests, the latter never went up higher than 0.65 mg/l.p.m. and never went down below 0.08 mg/l.p.m. The alkalinity of pure glacier ice taken off the Garabashi Glacier tongue equaled 0.08-0.09 mg/l.p.m., while the alkalinity of water taken directly from under the Garabashi Glacier tongue made up 0.45 mg/l.p.m. It is, therefore, safe to maintain, that when the water alkalinity in the river bed equals approximately 0.08-0.10 mg/l.p.m. the ground waters are completely "locked off" and the run-off consists only of melt and sub-glacial waters. The volume of the last mentioned as compared with the volume of melt waters is considerably small and despite their high content of alkalinity, they do not constitute a significant factor in the daily peak of liquid and salty runoff.

When the alkalinity reaches up to 0.45 mg/l.p.m. and higher, the role of melt waters in the runoff dwindles to zero and is defined only by the ground and sub-glacial waters. In the interval between 0.45 to 0.08 mg/l.p.m. the runoff is mixed.

The above stated is confirmed by the curve $A = f(Q)$ (Fig. 1), built on the data obtained from the analysis made on the majority of series. The curve has outlines approaching the hyperbola, its branches, however, becoming the parallel Y-axis beginning with the values of $A = 0.45$ mg/l.p.m. and $A = 0.08$ mg/l.p.m. The bottom

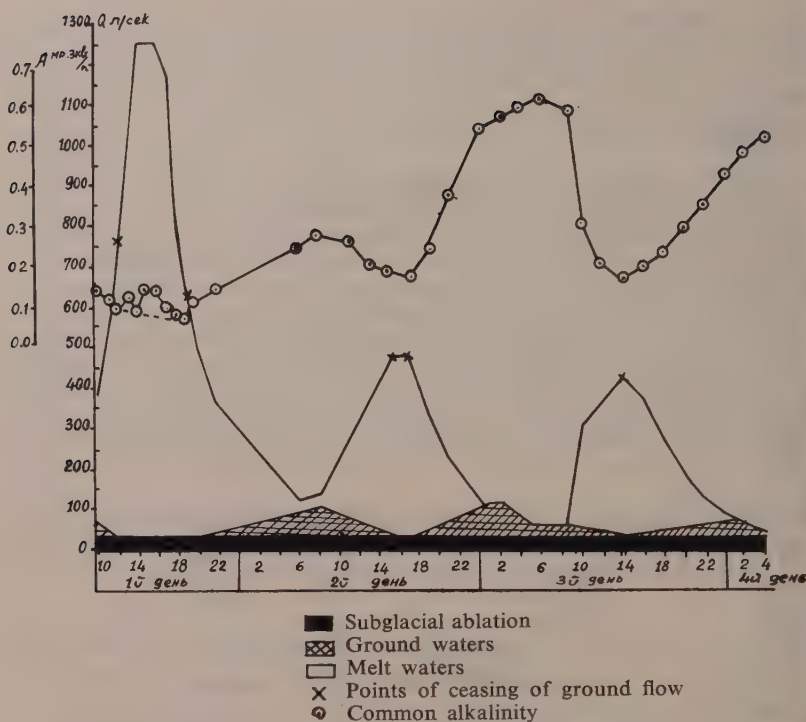


Fig. 2 — Disintegration of constituents of the flow.

point corresponds to the moment, when the alkalinity, upon reaching its minimal value, ceased to change, despite the rapidly rising water consumption, i.e. the runoff is then only defined by the melt waters possessing a relatively constant chemical composition.

The top breaking point beginning with $A = 0.45$ mg/l.p.m. corresponds to the minimum constant water discharge and the rapidly increasing alkalinity. The water discharge is determined by the controlled subglacial runoff (equalling about 30 l/sec as may be seen from the same curve) and by the ground waters, the mineralization of which increases in proportion to the depth of the releasing strata.

G. V. Tzytzarin⁽³⁾ recommends to adopt for plain rivers the method of separating the underground runoff component on the basis of hydrochemical data, which was accepted as the initial point for observations on the Garabashi River. Using the ratio between the salt composition and the water discharge on the Volga River, he separates the sections of various types of feeding on the curve. These sections are analogous to those mentioned above; section I, corresponding, as maintained by G. V. Tzytzarin, to cessation of the ground water runoff and increase of the surface runoff to our curve $A = f(Q)$, expressed by the middle part of the hyperbola ($A =$ from 0.45 to 0.08), Section II transmitting into a straight abscissa, parallel to the axis, in the event of hyperbolic dependence, and characterising the zone of high flood, in our case also corresponds to the maximum daily discharge (at $A = 0.08$) and finally, the appearance of ground water component and the cessation of surface runoff (which are the III and

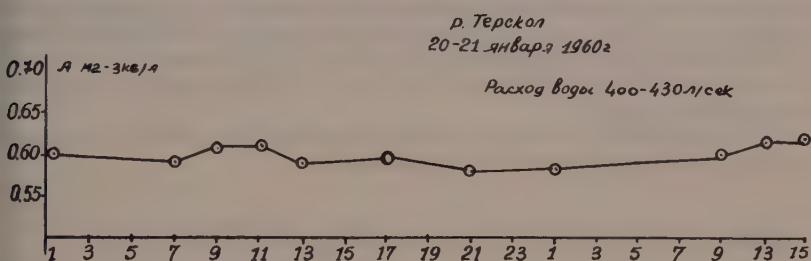
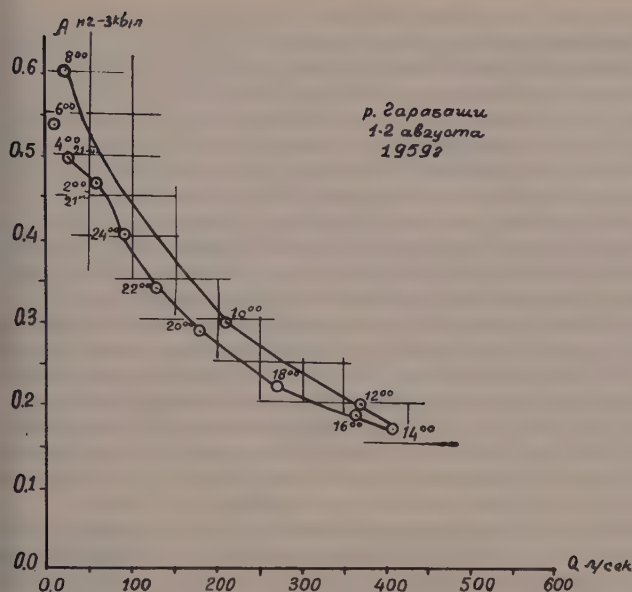


Fig. 3 — Summer and winter types of distribution of common alkalinity in the rivers of the Elbrus.

V sections according to G.V. Tzytzarin) are expressed by the top branch of the hyperbola $A = f(Q)$, which is parallel or almost parallel to the Y-axis (at $A \geq 0.45$). Thus, the nature of interrelation between the runoff and the chemical elements on a plain river is analogous to same on a river of snow-glacier feeding although it differs in time scale.

Fig. 2 shows that the inertia of changing the chemical composition depending on runoff is practically absent (even in one hour interval the linkage comes out quite satisfactory), therefore, the method described above may be applied to rivers with different sources.

In conclusion, we shall cite here the results of hydrographic investigations based on hydrochemical data of the daily runoff on the Garibashi River. Sampling for general alkalinity was carried out on the river at intervals of no more than two hours (during high flood, the interval was one hour). The postglacial runoff, as has already been mentioned, was kept uniform and, during the entire period of adjustment, was equal to 30 l/sec., the total alkalinity being over 0.45 mg/cpm/l. On passing the

high-flood, the ground water component comes down to zero, which is witnessed by the fact that the water alkalinity in the river bed approaches the alkalinity of glacial ice (0.10 mg/epm/l). During mild floods (2,3 day observation, Fig. 2) the role of post-glacial waters in the runoff remains quite noticeable and the alkalinity does not drop below 0.15-0.17 mg/l. On the other hand, when the flood is high (1 day observation, Fig. 2), already by 12 o'clock noontime $A = 0.10$ mg/l and remains close to this figure up to 9 P.M. As may be seen from the curve on Fig. 1, such alkalinity corresponds to complete cessation of ground water runoff. The points of cessation and starting ground runoff are shown on Fig. 2 by crosses. The increase of alkalinity at 3 and 4 P.M. of the first day of observation is explained by the influence of rapidly augmenting hydrostatic pressure in the bed, which caused the pressing out of considerable quantities of mineralized ground waters from the deep strata. The moments of discontinuing water inflow from the glacier surface (including the gradual emptying of the entire river bed network) are marked by transition values of alkalinity through 0.45 mg/epm/l. On the second and third day of observation this took place at about midnight. The surface runoff, as may be seen from the chart does not appear until about 9 A.M., when the active inflow of melt waters begins and causes a drastic decrease of alkalinity. The night after the first day of observation, when the minimum temperature in the region of alimentation was exceptionally high (+ 5.8) and thawing did not stop during the whole night, constitutes an exception from the rule.

Having determined the volume of subglacial runoff on the basis of alkalinity, and the moments of starting and finishing of the ground runoff by the curve on Fig. 1, we can point out in the hydrograph the value of all three components, as it is done on Fig. 2. In the winter period, when there is no surface runoff, the discharge and alkalinity remain practically constant. The alkalinity on the Tereskal River, approaching the conditions on the Garabashi River, fluctuates somewhere near 0.60 mg/epm/l, which approximately corresponds to the alkalinity of ground and subglacial waters (Fig.3).

Thus, the method of hydrograph investigation on rivers fed by snow glaciers embraces the following:

1. Choice of the salt component element of the given river, which is most easily defined and well reflects the runoff fluctuations.

No. of test	Date hour	Discharge	Ionic Content mg/l							
			Cl'	SO ₄ '	HCO ₃ '	Mg	Ca	Si	ions	pH
1	4/VII 4 P.M.	0.68	14.20	0.77	14.63	3.04	2.004	0.001	34.64	5.9
2	11/VII 1.30 P.M.	0.35	3.55	0.86	12.20	0.61	2.004	0.002	19.22	6.4
3	18/VII 1.30 P.M.	1.09	3.55	0.48	9.76	1.22	1.002	0.002	16.01	6.0
4	27/VII 13.00	0.40	10.64	0.48	17.08	1.82	1.002	0.004	31.02	6.2
5	27/VII 1 P.M.	ice	3.55	0.42	9.76	1.22	1.002	0.0005	15.95	5.8
6	19/VII 3 P.M.	sub-glacier	9.15	1.44	7.32	0.61	1.002	0.01	19.52	6.0

2. A series of definitions made on this element, the sampling time being so arranged, that all phases of flood during 24 hours should be well covered by observations. Observations should continue without interruption for a period of 2-3 days.

3. Draw up a curve $A = f(Q)$ on the basis of these observations and mark the sections, where there is no surface runoff, mixed runoff, or ground runoff. The points of altered regimes on the curve $A = f(Q)$ correspond to the values of A , when the curve branch lines become parallel to Y -axis.

4. The volume of the subglacial runoff is determined by the content of the defined element in the water taken directly from under the glacier tongue (it is best done during the absence of surface runoff, otherwise the subglacial runoff may prove exaggerated) and the curve $A = f(Q)$.

5. On the chart of the runoff course point out the volume of subglacial runoff (as constant) and mark the moments, when the ground and surface runoff stop. By connecting the corresponding points, the separate hydrograph components may be found.

In conclusion we cite below the table showing the ion composition of the Garabashi River during July-August 1959.

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THE ABLATION SEASON ON GILMAN GLACIER NORTHERN ELLESMERE ISLAND

G. HATTERSLEY-SMITH ^[1], J.R. LOTZ ^[2] and R.B. SAGAR ^[3]

RÉSUMÉ

Au cours des étés de 1957 et de 1958, des études intensives en glaciologie et en météorologie glaciaire furent effectuées sur le Gilman Glacier de même que sur la calotte glaciaire attenante sise en Ellesmere Island septentrional. Des études moins intensives furent aussi effectuées en 1959. Des différences significatives des conditions météorologiques glaciaires se sont produites pendant les trois années, ces différences se sont reflétées dans les totaux d'ablation mesurés sur la partie plus basse du glacier. Les totaux quotidiens d'ablation ont indiqué une relation très précise avec la température maximum quotidienne. On en a conclu que le régime du Gilman Glacier est négatif au temps actuel. Le terminus du glacier ne s'est pas retiré récemment, bien que le glacier est en train de s'atténuer dans les régions plus basses. Des traits au bord du Gilman Glacier, la position au temps actuel d'un glacier tributaire et des masses locales de glace, indiquent qu'il y a une retraite glaciaire bornée dont la prolongation de la saison d'ablation, et les plus hautes températures d'été sont les causes les plus probables.

SUMMARY

During the summers of 1957 and 1958, detailed glaciological and glacial-meteorological studies were made on Gilman Glacier and the adjoining ice cap in northern Ellesmere Island. Limited data were also obtained from the glacier in 1959. Significant differences in the glacial-meteorological conditions occurred in the three years; these differences were reflected in the amounts of ablation measured on the lower part of the glacier. Daily amounts of ablation showed a close correlation with daily maximum temperatures. It was concluded that the regime of Gilman Glacier is negative at the present time. The terminus of the glacier is probably thinning in its lower reaches. Marginal features of Gilman Glacier and the present status of a tributary glacier and of local ice masses provide evidence for limited glacial recession, of which lengthening of the ablation season and higher summer temperatures are the most likely causes.

INTRODUCTION

The work described in this paper was carried out as part of the glaciological and glacial-meteorological programme on Operation «Hazen», the Canadian I.G.Y.-I.G.C. expedition to the Lake Hazen area of northern Ellesmere Island, 1957-59, organized by the Defence Research Board.

Gilman Glacier, where the work was concentrated, drains the highland ice of the mountainous area north of the lake (Fig. 1); the névés at its head rise to an elevation of 2000 m. in a series of broad undulations which reflect an irregular subglacial surface. (1) The lower part of the glacier, flowing from north-west to south-east in a well-defined valley, is about 20 km. long and 4-5 km. wide; it is joined by three tributaries on the north-east side and one on the south-west side (Fig. 2). This part of the glacier falls from an altitude of 1200 m. to an altitude of 310 m. at the foot of its terminal ice cliff or ramp.

[1] Geophysics Section, Defence Research Board, Ottawa.

[2] Geophysics Section, Defence Research Board, Ottawa; formerly employed by the Department of Geography, McGill University, Montreal.

[3] Department of Geography, McGill University, Montreal.

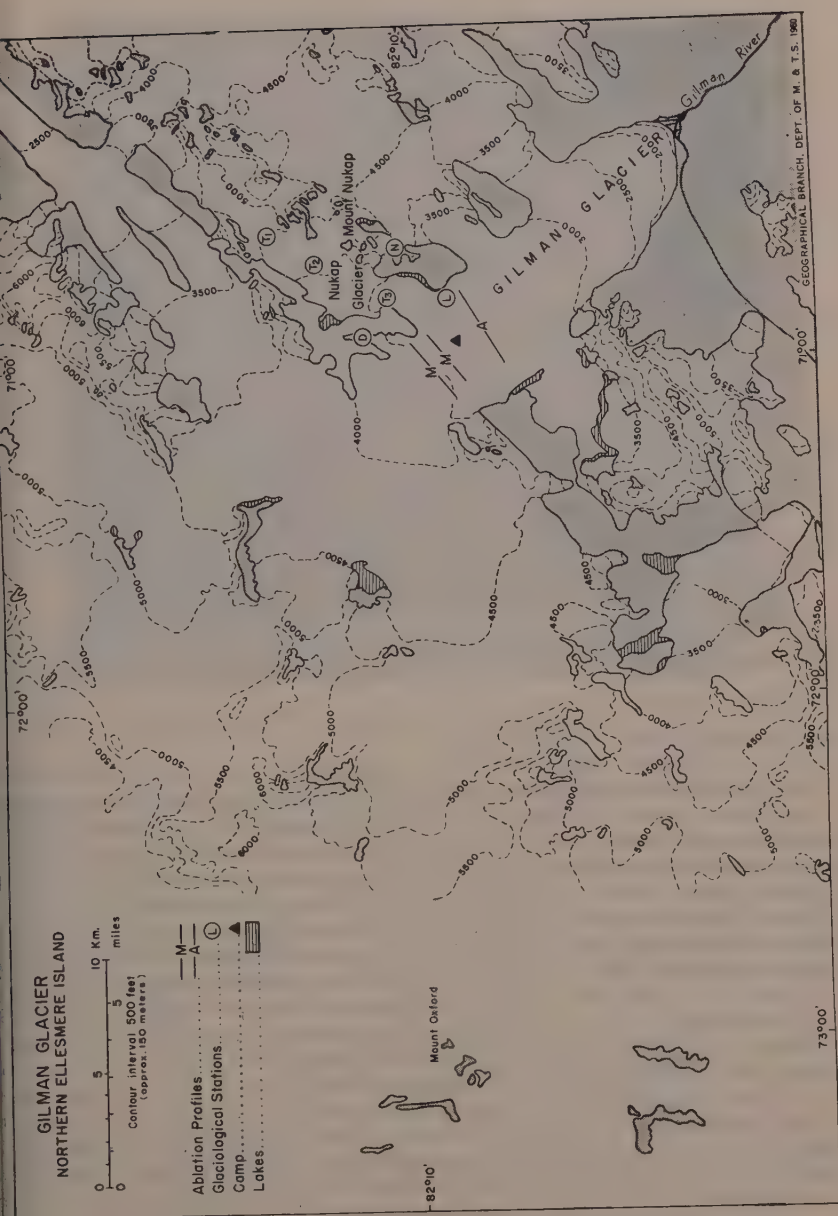


Fig. 1 — Map of Gilman Glacier, northern Ellesmere Island. (Prepared by Geographical Branch, Department of Mines and Technical Surveys: from lines from surveys by K.C. Arnold on Operation «Hazen»).



Fig. 2 — Gilman Glacier from the south-east, showing 1957-58 and 1959 camps (Δ). Photo from 9,000 feet (2740 m.) by R.C.A.F., 21 September 1956.

GLACIAL-METEOROLOGICAL INVESTIGATIONS

A programme of synoptic and micro-meteorological observations was carried out on Gilman Glacier to complement the glaciological work. The meteorological equipment was installed at a camp situated at an elevation of 1037 m., about 2 and 3 km. from the eastern and western edges of the glacier respectively (Figs. 1 and 2). Standard synoptic observations at 0800, 1400, and 2000, a micro-meteorological study of the air in the first ten metres above the glacier surface, and radiation studies were carried out here from early May until early August in 1957 and 1958. From early June to mid-August, 1959, a more limited programme of observations was carried out at a camp on the gravel flats close to the terminus of the glacier (Fig. 3).

Examination of synoptic weather maps, prepared by the Canadian Department of Transport arctic forecasting team at Edmonton, revealed that three broad weather groups affected northern Ellesmere Island successively in each of the three summers, 1957-59; they were related to dominance by high pressure systems in the early season, to weak cyclonic activity during the melt period, and to a conflict between high and low pressure systems in the late summer. In general, pressure gradients were weak and frontal systems very rare. Barometric pressure at the glacier station seldom varied by more than 2 or 3 mb. from day to day, and was essentially steady.

There are no long-term climatological records for northern Ellesmere Island. Records for Alert, situated 130 km. north-east of Gilman Glacier, date from 1950, but are of only limited value for comparative purposes, as the station lies on the shore of the Arctic Ocean, whereas Gilman Glacier lies far inland. Data from Alert indicate that greater positive temperature deviations from the mean occurred during the months of June and July 1957 (1.2°F) than in the same months of 1958 (0.4°F).

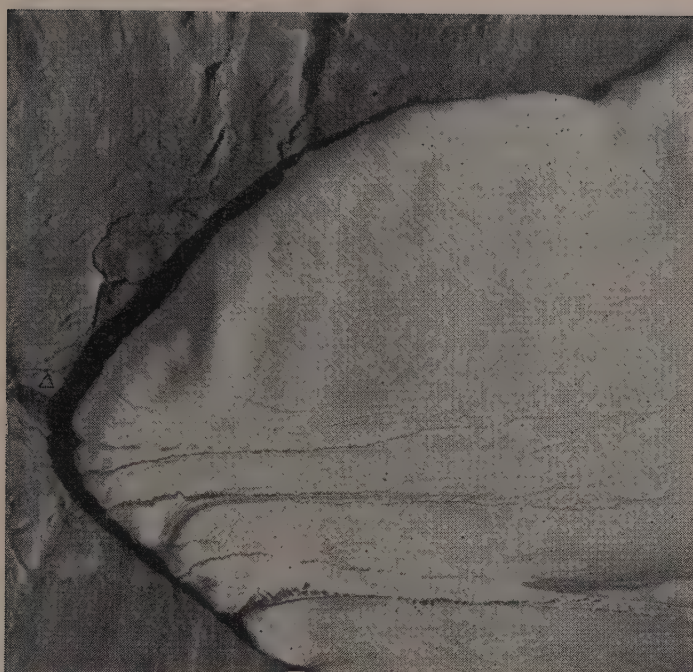


Fig. 3 — Terminus of Gilman Glacier from the east showing 1959 camp (Δ). Photo from 15,000 feet (4570 m.) by R.C.A.F., 2 August 1958.

While the same months of 1959 showed a negative temperature departure of 2.0°F. precipitation figures for Alert reveal deficiencies from the mean of 66 per cent in the 1956-57 budget year, and 36 per cent in 1957-58. The 1958-59 budget year figure of 100.0 cm. was very close to the mean for the 1950-59 period. Pit studies on the névés at 2000 m. at the head of Gilman Glacier did not indicate a significant decrease in budget year precipitation since 1950. ⁽²⁾

On Gilman Glacier, the early season until mid-June was characterized by long periods of clear skies and high sunshine totals. The following table shows the mean cloud cover between the hours of 0800 and 2200, and the mean sunshine total for the period 18 May to 20 June, 1957 and 1958.

	1957	1958
Mean cloud cover	3.1/10ths	5.0/10ths
Mean sunshine	86 %	71 %

Marked upsurges of mean daily temperatures occurred during early June; the temperature rose 12°F between the 8th and 13th in 1957. In 1958, this upsurge was

even more striking. On 6 June in that year, the mean daily temperature was 11.2°F while on 9 June it was 31.1°F. In 1959, the mean daily temperature at the glacier terminus rose abruptly on 7 June by 7°F.

After the onset of the melt season in mid-June, the period up to the end of July was characterized by high mean cloud cover, little sunshine, and relatively frequent precipitation in the form of snow, rarely in the form of rain. These conditions were more marked in 1957 than in 1958, as the following table shows for the period 21 June to 31 July in the two years.

	1957	1958
Mean cloud cover	6.0/10ths	7.7/10ths
Mean sunshine	13.3 hrs. daily	8.9 hrs. daily
Days of measurable precipitation	14	20

After the end of July similarities between the 1957 and 1958 seasons were less marked. In 1957, early August was characterized by a high mean cloudiness (8.4/10ths) and falls of rain and snow totalling 0.75 cm. In 1958, the first eight days of August had no rain or snow, and a mean daily cloudiness of 2.5/10ths. At the glacier terminus it would appear that the weather in August 1959 was intermediate between the conditions of the two previous years.

Other meteorological parameters can be summarized briefly from the available data. The sunshine total, expressed as a percentage of the possible total, was 65 per cent in 1957, 50 per cent in 1958, and 60 per cent in 1959. The latter figure, which was for the camp at the glacier terminus, was probably 10-15 per cent higher than the figure for the camp on the glacier, according to observations of cloud ceiling and distribution. As might be expected, solar radiation varied more or less directly, and mean cloudiness inversely, as the sunshine total. The air above the surface of the glacier during 1957 and 1958 was seldom saturated, and mean relative humidity for most days was about 80 per cent. The prevailing wind in both years was from the north-west, or down-glacier; 50-55 per cent of the observations showed winds

	1957	1958
Mean daily temperature	35.6°F	34.6°F
Mean maximum daily temperature	39.3°F	36.9°F
Days with maximum temperature over 40°F	29	7

from this direction. Calms were recorded at 20 per cent of all observations during 1957 and 1958—a surprisingly large percentage which reflects the slow, almost stagnant circulation of the High Arctic during the summer months. Mean wind speeds in 1958 were somewhat higher than in 1957, but in both years an increase of wind velocity from the 10 cm. to the 10 m. level was noted at over 50 per cent of the observations, characterizing a normal wind-flow pattern. Winds at all levels tended to be light, and seldom exceeded 4.5m./sec.

In spite of the later upsurge of temperature, the 1957 season was much warmer than the 1958 (Fig. 4). The last table compares mean daily and mean maximum daily temperatures for the period 20 June to 8 August in 1957 and 1958.

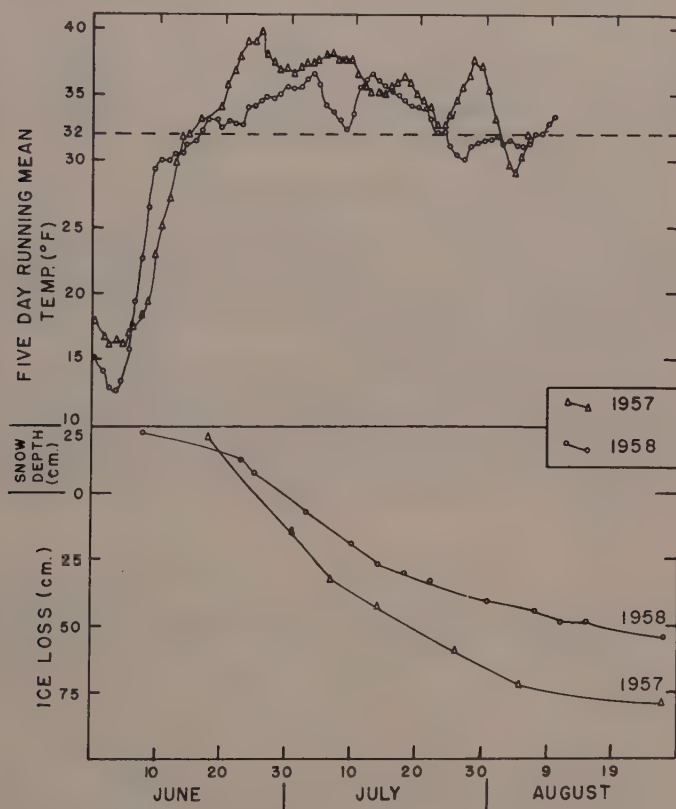


Fig. 4 — Five-day running mean temperatures at camp and ablation at «M» stakes on Gilman Glacier, 1957-58.

ABLATION ON GILMAN GLACIER

Lowering of the snow surface at the glacier camp through ablation, accompanied by density increases in the snow and refreezing of melt water at the ice surface, started on 12 June in 1957 and on 10 June in 1958. In 1957, all the snow had melted by 26 June, when ablation of the ice began; in 1958, ablation of the ice began on 28 June. From

28 June to 24 July, 1958, mean daily ice ablation was 1.2 cm.; from 26 June to 31 July, 1957, mean daily ice loss was 2.1 cm. The correlation between the rate of ice ablation and daily maximum temperature pointed out by Ahlmann (3) was found to apply to Gilman Glacier during 1957 and 1958. The following table shows typical amounts of daily ice ablation in early July; the values are the means of measurements at three stakes at the glacier camp.

Year	Date	Maxm. temp. (°F)	Ice ablation (cm.)
1957	4 July	41.4	5.3
	8 July	41.8	4.9
1958	6 July	44.8	3.4
	13 July	41.8	2.0

The table shows that, on days of comparable maximum temperatures, the ice ablation was less in 1958 than in 1957. In both years the type of ice being melted was similar, and consisted of a moderately bubbly variety, with a mean crystal diameter of about 0.4 cm. and larger crystals up to about 2 cm. in diameter. Albedo values were judged to be about the same in both years.

In May 1957, a network of ten stakes M1-10 («M» in Fig. 1) was established in two rows across Gilman Glacier, 1 to 3 km. north of the camp; the mean elevation of these stakes was 1049 m. The following table shows the mean snow depth and water equivalent of the snow in spring, and the mean ablation across the glacier at these stakes during three seasons of observations, 1957-59. The water equivalent (W.E.) for ablated ice is given on the assumption that the mean density of the glacier ice was 0.9 g. cm.⁻³.

Year	Snow Depth (cm.)	W.E. of Snow (g.cm. ⁻²)	Ice Ablation (cm.)	W.E. of Ice (g.cm. ⁻²)	Net Ablation (g.cm. ⁻²)
1957	23	6.5	78.2	70.4	77
1958	22.5	6.5	54.9	49.4	56
1959	21.5	7.3	38.0*	34.2*	41.5*

(*) at four stakes only up to 15 August.

Fig. 4 shows the progress of the ablation at these stakes and the five-day running means of mean daily temperature at the glacier camp for the 1957 and 1958 seasons. Ablation was greatest in periods of high daily maximum temperature, as shown for example in the following table for early July 1957 and 1958 at the glacier camp.

	2-9 July	
	1957	1958
Mean Maxm. Temp. (°F)	40.6	39.3
Mean Ablation of Ice (cm.)	30.9	13.5

In 1958 ablation measurements over a longitudinal line of stakes, set above and below the camp, and over four transverse lines of stakes below the camp showed a roughly linear relationship between ice ablation and elevation, from the equilibrium line at 1200 m. to the 830 m. level (Fig. 5). The mean ablation of ice over the whole area in

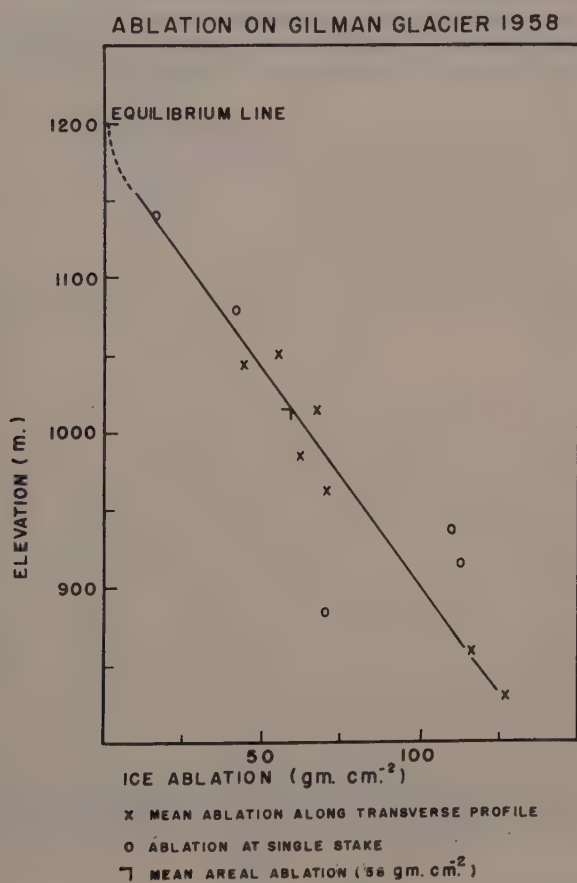


Fig. 5 — Ice ablation at various elevations on Gilman Glacier, 1958.

1958 was estimated at 58 g. cm.^{-2} . Below the 830 m. level few ablation measurements were made, but it appeared that in this area the relationship of ablation to elevation was non-linear. This was attributed to the increasing steepness of the glacier with its southerly aspect making available a larger amount of solar heat for melting the ice, and to the increasing proximity to bare rock near the snout of the glacier and consequent greater insolation effect. The mean ablation over this lower part of the glacier in 1958 was estimated at 145 g. cm.^{-2} of ice.

Owing to a preponderance of wind-blown rock material and to the insolation at bare rock surfaces, ablation is much greater near the margins of the glacier than away from the margins. In 1958, ablation measurements were made at three stakes—L1, L2 and L3 («L» in Fig. 1)—set 50, 60 and 50 m. from the eastern margin of Gilman Glacier near its junction with Nukap Glacier; the measurements are plotted in Fig. 6. The results showed a mean ice loss for 1958 of 165 g. cm.^{-2} , and an ice loss for 1959 at L1 (the other stakes had fallen down) of 137 g. cm.^{-2} up to 16 August. Fig. 6 also shows the mean ablation of ice at the transverse row of five stakes (S1-5) at mean elevation of 1015 m («A» in Fig. 1), and also the ablation at the stake S6, situated 800 m west of stakes L1-3. These results suggest that the ablation at S6 was very little affected by proximity to the margin of the glacier.

The results of detailed ablation measurements at the glacier terminus in 1959 have not yet been compiled. Owing to the great thickness of the glacier and a rate of movement of about 11 m./year ⁽⁴⁾, the frontal cliff or ramp is holding its position on

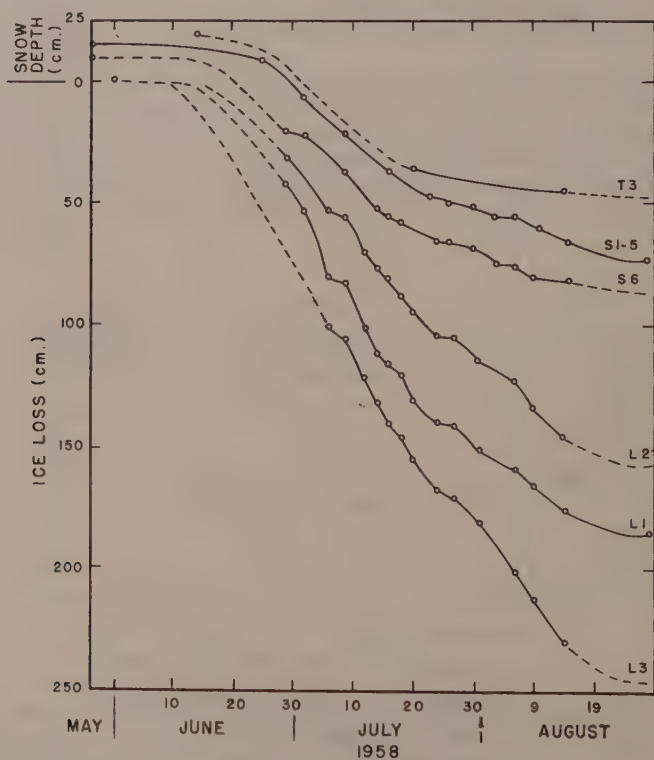


Fig. 6 — Ablation at stakes L1-3, S1-6 and T3 on Gilman Glacier, 1958.

anything, slightly advancing. The amount of calving from the ice cliff has only a minor effect on its position. The effect of bedrock ridges beneath the glacier is well-shown in the air photograph (Fig. 3); on the west side of the glacier the sub-surface ridge has impeded flow and caused excessive thinning on the down-glacier side.

A quantitative interpretation of the meteorological results for the three seasons is in progress. The amounts of ablation during 1957, 1958, and 1959 have been related to temperature in the following table, in which ablation and accumulated daily maximum temperatures* for 1958 are assigned unit values, and for 1957 and 1959 proportional values.

		1957	1958	1959
Accumulated daily maxm. temp. at glacier camp*		1.5	1.0	0.9
Ablation of ice	Poles M1-10 (1049 m.) («M» in Fig. 1)	1.6	1.0	0.75×
	Glacier camp (1037 m.)	1.4	1.0	0.85
	Poles S1-6 (1015 m.) («A» in Fig. 1)	1.3	1.0	1.0

× at four poles only.

ABLATION ON ICE DOME NORTH OF GILMAN GLACIER CAMP

The foot of the ice dome is situated 3 km. north of the Gilman Glacier camp. It rises steeply from a height of 1070 m. to its highest point at 1340 m. in a distance of 1.4 km. Except along the melt-water lake at the bottom of the south side, it is surrounded by ice-free land from whose contours it is safe to say that the dome has a maximum ice thickness of 50-100 m. (Fig. 7).

Ablation data on the ice dome are available from July 1957 to August 1959 (Fig. 8). Measurements were made at three stakes—D1, D2 and D3 («D» in Fig 1.) at elevations of 1340 m., 1305 m., and 1180 m.; D1 was situated at the summit (Fig. 7.) The equilibrium line on Gilman Glacier lay at an elevation of about 1240 m. in 1957 and at about 1200 m. in 1958. In 1957, local climatic conditions on the ice dome gave an equilibrium line appreciably higher than its summit, where there were 10 cm. of ice ablation; in 1958, the equilibrium line was close to the summit level, and 3 cm. of superimposed ice were formed. The ablation of 40 cm. of ice on the summit in 1959 may have been due to a very low spring snow cover, for on Gilman Glacier itself the summer appeared from the ablation records to have been rather cooler than in 1958. Over the three summers the total ice loss due to melting at the summit amounted to 56 g.cm.⁻².

* The sum of all degrees above 32.0°F of the daily maximum temperatures.

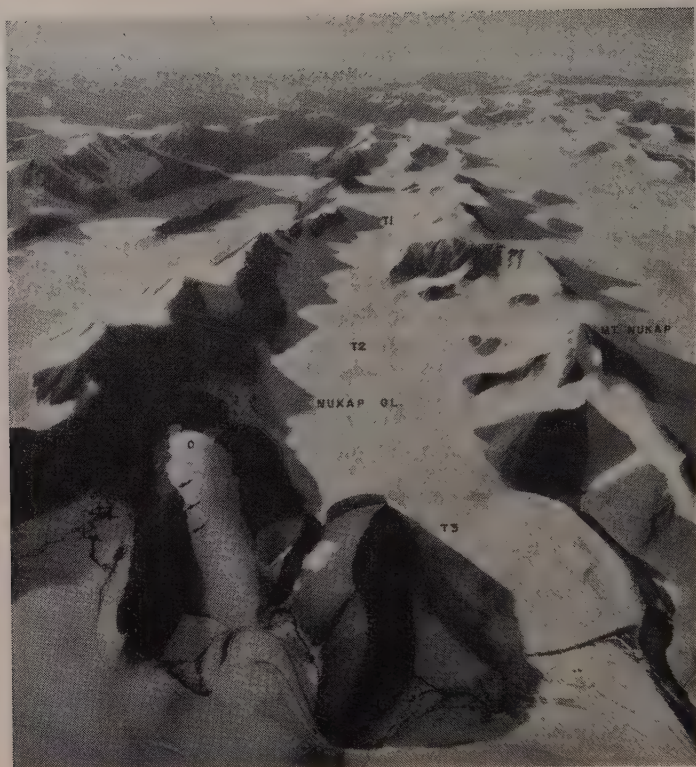


Fig. 7 — Nukap Glacier, Mount Nukap and ice dome («D») from the southwest showing glaciological stations (L, N, T1, T2, T3), and marginal cairns (x). Photo from 15,000 feet (4750 m.) by R.C.A.F., 2 August 1958.

A core at D1 showed alternating layers of clear ice, with a few bubbles, and milky ice, with numerous bubbles, to a depth of 1.6 m. There were several layers of dirt associated with small cryocinite holes. An ice core from a depth of 60 cm. showed a distinctly granular texture with no intergrowth of the crystals, which ranged in diameter from 0.5 to 1.25 cm.; the density was 0.79 g. cm.^{-3} . The crystal structure and density was similar to that of the superimposed ice being formed near the equilibrium line on Gilman Glacier at the present time. The present period of excess ablation at the summit appears to be uncovering ice formed in a recent slightly colder period, and not relic ice in which the crystal texture would be intergrown and the density about 0.9 g.cm.^{-3} .

On the west side the edge of the ice dome is retreating rather rapidly, as might be expected from the overall ablation. The retreat is exposing patterned ground (stone stripes) and a few dried remnants of vegetation. Silt and sand, which form the surface dirt layer of the melted ice, lightly mantles the drift material for a distance of up to 50 m. from the ice edge. The following table shows the amount of retreat from the cairns whose positions are shown in Fig. 7.

The increased melting in summer during the last twenty years, deduced from pit studies in the firn at an elevation of 1800 m. near Mount Oxford (²), has probably

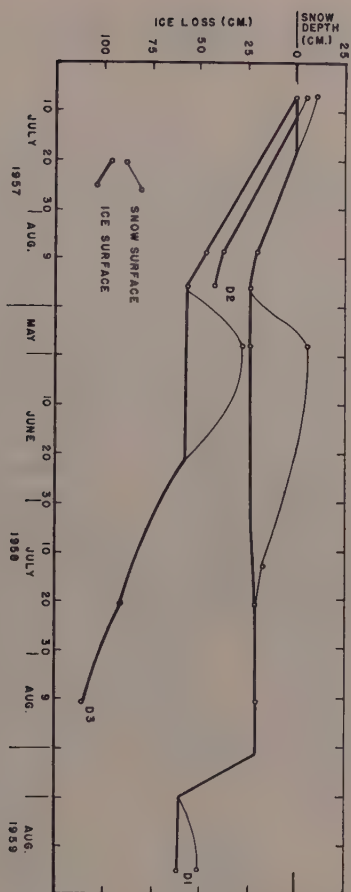


fig. 8 — Accumulation and ablation on ice dome, 1957-59.

Cairn	Elevation (m.)	Retreat (m.)	
		21 July to 10 Aug. 1958	21 July 1958 to 16 Aug. 1959
1	c. 1220	2.75	5.03
2	c. 1190	1.5	(not measured)
3	c. 1165	3.0	14.1
4	c. 1150	2.25	8.18

been sufficient to put the level of the equilibrium line above the summit level of the ice dome in most years during this period.

NUKAP GLACIER

Nukap Glacier flows down from the ice divide separating it from a glacier flowing steeply towards Clements Markham Inlet on the north coast of the island; it joins Gilman Glacier 3 km. east of the glacier camp (Figs. 1 and 7). The results of 1958 pit studies at the top of the glacier and in the middle and lower sections, at stations T1, T2 and T3 respectively, are shown in Fig. 9. The pit profile at station

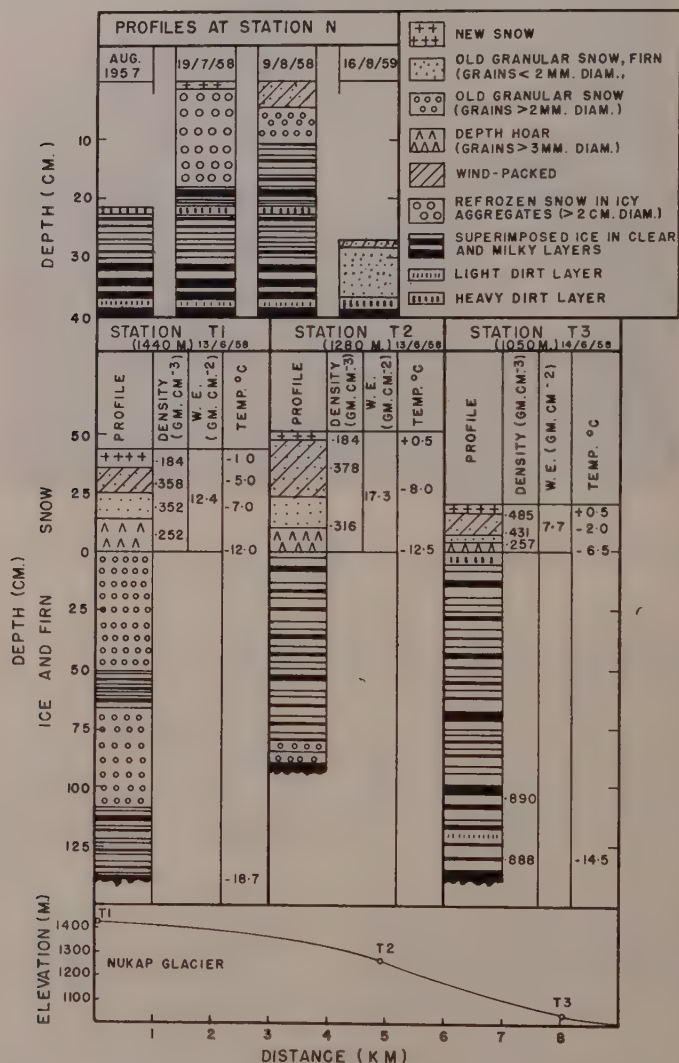


Fig. 9 — Accumulation and ablation on Nukap Glacier and on Mount Nukap 1957-59.

T1 (1440 m.) showed mainly firn, with grains up to 3 mm., and lesser thicknesses of banded, milky and clear ice layers beneath a snow cover of 43 cm. At this elevation the freezing temperature is probably reached in at least the top metre of the firn and snow pack each summer. This means that percolating melt water may penetrate and refreeze in the accumulated firn of several years. No annual interpretation of the stratigraphy in the 1.8 m. pit was therefore possible.

At station T2 (1280 m.) a very minor amount of firn was interfingered with banded, superimposed ice down to the bottom of the pit at 1.45 m. This pit was a little above the equilibrium line, while station T3 at 1050 m. was considerably below it, for the ice ablation here during the 1958 summer amounted to 40.5 g.cm.^{-2} up to 14 August. This compares with a mean ablation of 44 g.cm.^{-2} at the stakes M1-10 situated at a similar elevation. The accumulation and ablation areas of Nukap Glacier were roughly equal in 1958. With a maximum ice ablation of not less than 40 g.cm.^{-2} a maximum accumulation unlikely to exceed 18 g.cm.^{-2} , it seems clear that this glacier and experienced an appreciable deficit in 1958.

The lower part of the glacier appears to be constricted in its flow by the two spurs at the 1150-m. level, and by a sub-glacial ridge which probably joins them. Thinning of the ice over this ridge may cause a reduced flow to the lower part of the glacier. The moraine on the surface of the glacier near its junction with Gilman Glacier indicates recent wastage over a fairly long period. Much of the rock material has been transported for a considerable distance, since large blocks of limestone with Permo-Carboniferous fossils are not known to occur nearer than 15 km. from the site. In order to account for the very large size of some of the rocks, it is suggested that the moraine has mainly been carried there by surface glacial streams of great power in the summer. Near the nunataks on either side of the junction, Nukap Glacier is receding slightly. The marginal river on the south-east side of Nukap Glacier is tending to lower its bed as it cuts into the side of the glacier. This has resulted in the formation of ice gorges and the gradual destruction of the few remaining ice ramps connecting the glacier to the nunatak. On both sides, the recession is marked by the gradual encroachment of the marginal lakes towards the glacier and, on the south-east side, by the development of ice-cored gravel cones and ridges of moraine (Fig. 10).

ICE COVER ON MOUNT NUKAP

Mount Nukap rises steeply to 1782 m. from glaciers to the north-west and south-east, and less steeply from the nunatak to the south. Except on its south and east sides, the peak is largely ice-covered. The ice apron on the south side puts out a small glacier tongue which terminates 200 m. from Nukap Glacier at an elevation of 1040 m. Southward along the ridge the ice apron thins out, and finally disappears at an elevation of 1130 m. on the bare ground of the nunatak (Fig. 7.)

On 19 July 1958, an ablation stake was set at an elevation of 1370 m. in the ice apron on the south ridge (station «N» in Fig. 1). A pit here showed granular, refrozen and recrystallized snow in 2-5 cm. aggregates, overlying 3.5 cm. of superimposed ice formed during the summer. Beneath the snow, a dirt layer with cryoconite holes represented the 1957 summer surface. An ice core showed alternating layers of milky and clear layers down to a depth of 113 cm. below the snow surface, with fine layers of windblown dust at depths of 28 and 103 cm. (Fig. 9). The larger ice crystals ranged from 0.5 to 1 cm. in diameter. There was no firn in the core. It was considered that the ice had accumulated as superimposed ice, or as water-soaked and refrozen firn. On 9 August 1958, at the end of the ablation period at this elevation, 7.5 cm. of superimposed ice remained beneath 10.5 cm. of recent snow, partly recrystallized after watersoaking and partly wind-packed. On 16 August 1959, measurements at the stake

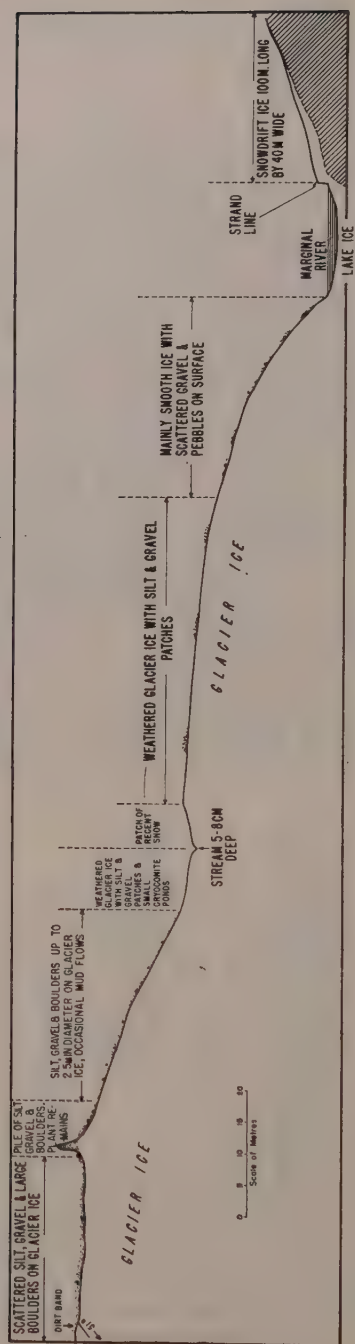


Fig. 10 — Profile across south-east margin of Nukap Glacier near junction with Gilman Glacier, July 1958.

showed ice ablation for the 1959 summer of 19.5 cm., including 11 cm. of superimposed ice formed during the 1958 summer; there were 10.5 cm. of recent snow at this date. The snow cover on this exposed ridge is no doubt very variable from year to year; it was probably light in the 1959 spring, which would account for the ablation of ice that occurred. In some years, as in 1958, there may be a small accumulation of superimposed ice. Over a few years it seems certain that there is net ablation, in contrast to conditions at the same elevation on Nukap Glacier where the snow accumulation is greater. The ice apron of Mount Nukap, where it adjoins bare ground, is retreating by several metres every year.

The effect of the present recession and thinning of small ice masses is best shown by a development of snowdrift ice at 1160 m, situated 400 m from the south edge of the Mount Nukap ice apron. On 19 July 1958, an ablation pole was set in the snowdrift ice, and an ice core taken. The ice core to a depth of 88 cm showed mainly clear ice with a few bubbles up to 1 mm in diameter; the crystals ranged in diameter from 0.25 to 1 cm. There were two 2-cm bands of milky and very bubbly ice at depths of 30 and 40 cm, and two 2-cm layers of ice containing windblown dirt at depths of 50 and 65 cm. Between 19 July and 9 August, 44 cm of ice ablation was recorded, and near the centre of the ice mass bare rocks, over an area of about 1 m² on 19 July, covered an area of about 70 m² on 9 August. The ice edge had receded by up to 1 m all around. By 16 August 1959 the ice, which was 50 to 100 cm thick in July 1958, had disappeared from an area of about 700 m², and remained only in an area of about 100 m², in a hollow where it was most protected from the sun.

It seems safe to conclude that a gradual thinning and disappearance of ice has been taking place on this nunatak in recent decades.

ABLATION ON GILMAN GLACIER AND THE REGIME

At the present time, the accumulation between an elevation of about 1450 m. and the highest part of the ice cap at 2000 m. is by firn formation; between an elevation of about 1280 m. and 1450 m. interfingering of firn and superimposed ice takes place. Accumulation exclusively by formation of superimposed ice probably only occurs in a fairly narrow belt between the equilibrium line at about 1200 m. and an elevation of about 1280 m. (2). The value of accumulation - ablation at the equilibrium line is approximately 7.5 g.cm.⁻². Measurements of accumulation and ablation indicated a deficit for Gilman Glacier of about 33 million m³. of water in the 1957-8 budget year, which was about 60 per cent of the total accumulation for that year, and a larger deficit in 1956-7; in 1958-9 the glacier appeared to have been more nearly in a state of balance (5). It seems clear that the regime of the glacier at the present time is negative. But, unless a deficit of the order noted in 1958 prevailed over many decades, its effect on the area of Gilman Glacier would not be pronounced for two reasons. First, Gilman Glacier is tapping a vast reservoir of ice from the highest part of the ice cap, where small climatic changes have least effect. Secondly, the glacier rises very steeply from its terminus (Fig. 3) and reaches a thickness of 400 m at a distance of only 5 km therefrom, and a thickness of 600 m near the glacier camp (6); thus the amount of very cold ice which melts in a relatively short ablation season causes a scarcely significant thinning. In fact the glacier has certainly not receded in recent years, and may even have advanced slightly.

Nevertheless marginal features of Gilman Glacier and the present status of its tributary, Nukap Glacier, and of local ice masses do provide evidence of a limited glacial recession. This recession is related to the present negative regime of Gilman Glacier, of which lengthening of the ablation season and higher summer temperatures are the most likely causes.

ACKNOWLEDGEMENTS

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DISTRIBUTION OF THE SNOW ON THE GLACIERS OF THE ZAILIYSKY ALATAU

K.G. MAKAREWICH (U.S.S.R.)

SUMMARY

The observations on the distribution of snow on glaciers and the dynamics of snow masses began in July 1957 on the Tooyuksoo Glaciers of the Zailiysky Alatau (the North Tien Shan).

More than 3500 definitions of the density and the mass of snow cover in 150 points and more than 8000 measurings of its thickness had been made during 1957-1959.

All the principal types of the contemporary glaciers of the Zailiysky Alatau are concentrated in the Tooyuksoo group. They are approximately in the centre of the mountain-range with the maximal quantity of the atmospheric precipitations. The quantitative characteristics of the meteorological conditions are cited in the report.

The brief description on the points, profiles and routes of the observations of snow on the different glaciers is given below (tables). The formation of the snow cover on glaciers begins during the period from the end of August to the beginning of October. The gradual increase of the snow cover begins in November. The wind processes become more active at the same time due to the intensification of the general circulation of atmosphere, the cyclone invasions, the effect of the Central Asia anticyclone. The redistribution of snow masses on glaciers takes place due to the winds and orography of the basin.

At the end of winter (the beginning of June) the mass of snow cover increased on the upper parts of glaciers. At the same time on the lower parts the thickness and the mass of snow cover is quickly decreased independently from the increase of its density. It can be explained by the drainage of meltwater at the lower parts of the glaciers, while at the upper parts it densifies the snow and promotes its metamorphosis. Only at the end of June and in July snow melts quickly and remains only above the snow (firn) line.

The absolute maximum of the thickness, density and mass of snow is observed in May-June. The dynamics of snow on the special snow grounds and routes on the tongue and in the circus of the glacier, on the lateral and frontal moraines, on the slopes of different expositions is reported. The greatest quantity of snow is observed on the west side of the valley, where it is transported from the opposite slopes by west winds.

Data on the average thickness, density and mass of snow on different profiles of the Tooyuksoo Glaciers on the 1st of June 1958-59 are given in the report.

The map of distribution of snow over the glacier on the 1st of June 1959 is made. The areas with different snow masses is estimated by the planimetry. This is necessary for the prognosis of the flow of meltwater and their role in the nourishment of mountain rivers, for the irrigation, water-supply and generation of the hydro-energy.

The dependance between the mass of snow, the altitude, the distribution of atmospheric precipitations in the basin, the influence of snow and weather conditions on the height of the firnline in different years, on the mass balance of glaciers and on the length of the period of melting are examined. A comparison between the maximum mass of snow and the sum of precipitations at the same time leads to the result that 12-15 % of snow on a glacier is brought from the surrounding slopes and from the upper parts of the neighbouring basins.

RÉSUMÉ

Des observations furent commencées en Juin 1957 sur les glaciers de Toujouksou sur la crête nord du Tien-Chan — l'Ala-Tau transilique, de même que différents autres travaux glaciologiques sur la distribution de la couche de neige des glaciers et la dynamique des masses de neige. Ces travaux, poursuivies tout le long de l'année, sont encore en cours actuellement. Durant la période de la 2 A.G.I. — 1957-1959 — plus de 3500 estimations de densité et de réserve d'eau dans la neige ont été effectuées sur 150 différents points. L'épaisseur de la couche de neige a été mesurée plus de 800 fois en 400 points sur la surface des glaciers et des moraines.

Tous les principaux types actuels de glaciation de l'Ala-Tau transilique sont concentrés dans les glaciers de Toujouskou. Ces glaciers sont situés approximativement au centre de la chaîne, dans la région où les précipitations atmosphériques atteignent leur maximum. On trouve dans le rapport des caractéristiques quantitatives sur les conditions climatiques.

La couche de neige sur les glaciers a été évaluée au cours d'un mois — de la fin du mois d'août jusqu'au début du mois d'octobre.

La couche de neige augmente progressivement à partir du mois de novembre. C'est à cette époque que les vents deviennent plus actifs par suite de l'intensification générale de la circulation atmosphérique, de la pénétration des cyclones, de l'action de l'anticyclone sibérien. On note une redistribution des masses de neige sur le glacier liée aux vents et à l'orographie du bassin.

Vers la fin de l'hiver (début du mois de juin) les réserves d'eau dans la neige augmentent sur les secteurs supérieurs du glacier. Sur les secteurs inférieurs, malgré l'augmentation de la densité, la puissance de la masse de la couche de neige diminue intensivement. Ceci s'explique par le fait qu'une partie des eaux de fonte, dans la partie inférieure de la langue ruisselle, alors que sur les secteurs supérieurs du glacier, les eaux de fonte tassent la neige et contribuent à sa métamorphose. Ce n'est qu'à la fin du mois de juin et en juillet que la neige fond activement, demeurant intacte en amont de la ligne de névé.

Le maximum absolu de hauteur, densité et masse de la couche de neige est noté en mai-juin.

Dans le rapport on a considéré la dynamique de la couche de neige sur des aires de mesure spéciales, le long des itinéraires sur la langue et le cirque du glacier, ainsi que sur les moraines latérales et frontales et les versants ayant différentes expositions. La plus grande quantité de neige a été remarquée sur la face occidentale de la vallée par suite de la mutation sur ce côté de la neige des versants des contreforts opposés par les vents des aires Ouest.

On trouve dans le rapport des données concernant la hauteur, densité et réserve d'eau moyennes sur les différents profils des glaciers de Toujouskou jusqu'au 1 juin 1958-59. Des cartes isométriques de distribution des réserves d'eau jusqu'au 1 juin 1959 ont été dressées en se basant sur les observations de la couche de neige.

Par la planimétrie des surfaces ayant une humidité de la couche de neige diverse, il fut possible de déterminer le volume d'eau dans la neige — facteur d'une haute importance dans les pronostics de l'écoulement des eaux de fonte et la détermination de leur rôle dans l'alimentation des rivières utilisées pour l'irrigation et la distribution d'eau.

On note dans le rapport la mesure dans laquelle la réserve d'eau dans la couche de neige dépend de la hauteur absolue, de la chute et la distribution des précipitations atmosphériques dans le bassin, de la dépendance de la couche de neige et des conditions atmosphériques au niveau de la ligne de névé (par années) sur le bilan matériel des glaciers et sur la durée de la période d'ablation. On trouve aussi des données comparatives sur les réserves d'eau maximum dans la couche de neige et la somme des précipitations au moment de son apparition. Approximativement 12-15 % de la neige sur le glacier peut être considérée comme étant de la neige venue des versants situés aux alentours et soufflée des hauteurs des bassins voisins.

In July 1957, in compliance with the I.G.Y. programme, the expedition of the Geographical Department of the Academy of Science of the Kazakh S.S.R. commenced complex glaciological researches on the Tuyuksuysky glaciers that are situated in the upper part of the river Malaija Almaatinka.

Simultaneously researches also began with the snow cover. This work practically consisted of observations of thickness, density, distribution of the snow cover, measurements of water stores in the snow, the quantity of surplus snow that refills the glacier substance in the accumulation regions.

The observation of the distribution of the snow and its influence upon the mass balance of the glaciers will be briefly expounded in this report.

The Tuyuksuysky glaciers are bedded on the northern slope of the Zailiysky Alatau-the advanced ridge of the Tyanshan that elevates over 4,000 meters above the plains of the south-eastern Kazakstan which open to the north. In these are concentrated all the essential features of contemporary glaciers of the said ridge. Situated almost in the centre of the ridge, they lie into the region of the heaviest precipitation. Among nine Tuyuksuysky glaciers situated at the sources of Malaya Almaatinka.

The Zentralny Tuyuksuysky is considered the main glacier; it occupies the principal part of the valley. On the east of it, there is the glacier Igly Tuyuksu. On the west of the lower part of tongue of the main glacier is situated the glacier Molodeshny (Fig 1).

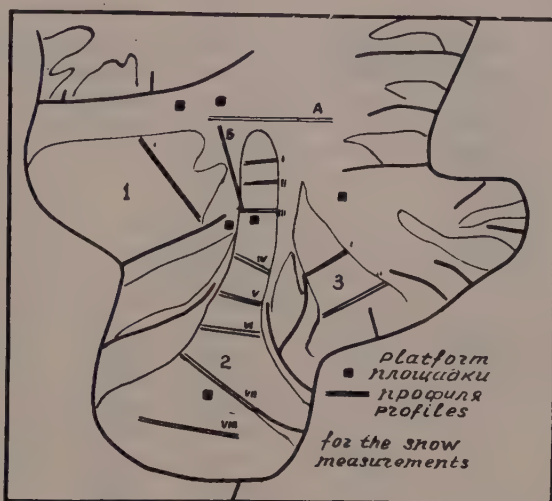


Fig. 1 — General scheme of glaciers of the river basin of Malaya Almaatinka in the Zailiysky Alatau. Glaciers Molodeshny (1), Zentralny (2), Igly Tuyuksu (3).

The average altitude of spurs, surrounding the mentioned glaciers, is 4,188 m (on the East) and 4,052 m (on the West), i.e. the first exceeds the latter by 136 m.

The main object of the observations of the snow cover during the period of the I. G. Y., was the glacier Zentralny, that extends from South to North 5,1 Km.

The climatic conditions in the glacial zone of the Zailiysky Alatau are very nearly similar to conditions of the polar regions.

Here are only four months (sometimes it is three months) during the year, that have a positive temperature. These months are: June, August and September. However, as a rule, it is colder in June than in September. Certain data about the air temperatures, cloudiness and precipitations in the zone of the Tuyuksuysky glaciers are given in the table 1.

Out of 30 months of uninterrupted observations, during 19-20 months the cloudiness was 6/10 to 9/10. In the summer months of 1957-59 the average cloudiness was about the same and reached 5/10. Almost three quarters of the year's precipitations fell in April and August.

The average monthly speed of wind seldom exceeds 3 m/sec. The highest speeds are noticed when cyclonal storms are present and reach 20-30 and even more m/sec. Windless days are almost absent, entirely because a mountain-valley-circulation regularly occurs.

The points of observations of the snow cover of the glaciers were located in different parts according to their respective and absolute altitude, relief, exposition, deviation, and character of the surface bedding.

Taking into consideration the existing danger in the mountains, it was necessary to avoid establishing observations in poorly accessible places of the basin. On the principal glaciers of the Tuyuksuysky group the mentioned points were established

Some characteristics of the meteorological conditions in the zone of the Tuysuksuysky glaciers during the IGY-IGC of 1957-1959.

[illegible]

TABLE 2

Characteristics of snow-measurement work in the region of the Tuyksuysky Glaciers

Profile Platform, route	Quantity of stakes	Average absolute altitude of profile, (m)	Average distance of profiles from end of open tongue (m)	Length of profile or dimensions of platform (m)	Deviation of surface along axis of glacier (degrees)	Remarks
GLACIER ZENTRALNY						
Profile						
1	10	3420	189	325	12	Profile laid on open surface of glacier.
2	11	3443	376	355	9	
3	11	3478	637	405	5	
4	13	3549	1047	485	11	
5	11	3624	1629	480	11	
6	8	3675	1957	525	7	
7	12	3730	2477	700	4	
8	8	3782	3100	585	6	
Platform	—	3480	650	50 × 100	5	on the tongue in the circus on the left
"	—	3750	2600	50 × 100	4	
"	—	3490	650	50 × 50	5	
Route	10	3415	below the end of the tongue 1200			side moraine on the frontal moraine.
GLACIER IGLY TUYUKSU						
Profile						
1	6	3567	650	400	9	
2	4	3617	1025	400	8	
Platform	—	3490	below the end of 50 × 50		5	
GLACIER MOLODESHNY						
Profile						
1	8	3511	179	580	9	on the slope of spur North of gla- cier.
Route "B"	8	3442	below the end of tongue 1200			
Platform	—	3450	—	50 × 50	15	
"	—	3450	—	50 × 50	15	

on such spots, where the researches of motion speed, thickness, hydrological regime of the glaciers, meteorological and radiation conditions, etc. were being made. Owing to this, almost all of them were bound topographically, so that it was possible to draw the "snow situation" on the available largescale maps with sufficient exactness the characteristics of snow observation points which are given in table 2.

The measurements of thickness and density of snow by means of a portable stake and weight cylinder, on the profile of motion and routes through moraine deposits of the glaciers were made during the winter periods of 1957-58-59 on the first day of each month. In the glacial zone solid precipitation fall during the whole year round. However, in the summer months, besides the solid precipitation, it also rains, and as a rule, by prolonged wet weather and strong cooling of the air, the rain devolves into snow, hail and sleet.

For the winter season the snow cover usually sets in during the period from the end of August until the beginning of October. In 1957 and in 1958 the snow cover became stable at the beginning of the second decade of September.

In 1959 the first snow-fall—the winter's forerunner—passed on September 11-th, but the condition of warm and clear weather with intensive solar radiation held on up to the 5 th of October. With the onset of a stable snow cover, the ablation period on the glaciers comes to an end and the ice melting ceases. Later, the snow cover experiences certain changes: its height and water content fluctuate, but after November their gradual growth is observed. Particularly the beginning of winter of 1958-59 was remarkable by a slow, but stable augmentation of the height of snow and an increase of watercontent (table 3).

TABLE 3

Average height h (cm) and watercontent W (mm) of the snow cover on the glacier Zentralny in the winter of 1958-59 (on the first of each month)

Months profiles	IX	X	XI	XII	I	II	III	VI	V	
I h	19	18	34	51	67	85	99	111	114	77
W	36	60	84	107	234	212	178	244	239	231
II h	16	15	30	45	55	69	85	94	104	69
W	35	42	63	81	181	165	153	197	208	207
III h	14	18	34	49	58	80	92	96	103	75
W	41	44	70	108	162	176	202	230	247	232
IV h	19	22	44	64	69	88	105	105	100	109
W	33	74	102	147	255	235	210	252	240	371
V h	26	22	38	54	60	83	102	103	118	134
W	63	84	102	124	210	183	214	206	246	387
VI h	39	25	40	54	—	67	73	87	142	134
W	62	78	90	107	120	134	182	226	355	389
VII h	38	27	47	64	—	88	93	119	163	149
W	66	88	108	128	160	220	270	286	407	462
VIIIh	—	—	—	68	—	86	85	111	145	137
W	—	—	—	135	180	224	230	233	333	411

As the observations have shown, the peak of density on all profiles occurs at the beginning of January. During December 1958 the altitude of the snow cover on the glacier Zentralny only slightly changed, but the water content had noticeably increased. The augmentation of the density is explained by the strong descending winds which especially in the months of December-January, actively blow the mass of snow and facilitate its compaction. In this period on the glacier Zentralny, three parts with the largest height of snow, density and water content are clearly recognized. The first part is located at the lower part of the tongue. The second part is situated in the region of the 4 th profile. The third stretches in the district of the 7 th profile. Here is noted the greatest accumulation of snow mass, this fact being connected with the increase of precipitation quantity and the transference of the snow from the 8 th profile and surrounding slopes, and also with the blowing over of snow from the basin of the river Bolshaya Almaatinka.

Such diffusion of the snow was preserved chiefly until the end of April—the beginning of May 1959, when the melting had begun. The largest of the average indications of thickness, density and watercontent fall in this period on 6 th and 7th profiles, where they reach respectively 142 and 163 cm, 0,25 and 0,25 gr/cm³, 355 and 407 mm. At the beginning of June, when the melting of snow increases, the height of the snow cover in all profiles diminishes in spite of a general increase of density. The water content in this period increases on 4-7 profiles only.

The melting of the snow on the latter profiles and its flowing down from them is most actively shown at the end of June and during July.

The absolute maximum of thickness, density and watercontent of the snow cover is noted in June (table 4).

TABLE 4

Absolute maximum of thickness, density and water content of the snow cover on the glacier Zentralny in 1958 and 1959.

Years	Profiles	1	2	3	4	5	6	7	8
Thickness of snowy cover (cm)	1958	207	228	198	300	245	220	223	—
		June							
	1959	100	128	128	128	180	155	171	165
		June		May		June		May	
Density of snow (gr/cm ³)	1958	0,38	0,38	0,34	0,37	0,36	0,35	0,32	—
		June							
	1959	0,32	0,32	0,32	0,40	0,32	0,34	0,33	0,33
		June							
Water content (mm)	1958	684	620	561	1110	676	665	682	—
		June							
	1959	210	224	240	427	576	420	521	465
		June							

The largest thickness of the snow in the cirque of the glacier Zentralny reached 4 meters. On the approaches of the pass Tuyuksu in a small basin that rises over the

glacier's cirque, the snow cover in April equalled 9 meters, which was conditioned by the accumulation of snow that is being carried over from the basin of the Left Talgar, and the fall of snow avalanches from the peak Pogrebitsky.

The observations on the snow measurement platforms that were made in ten-day periods give a full idea of the dynamics of thickness, density and humidity of the snow cover, both on the glacier Zentralny as well as on the moraines, the course of which is, in general, similar to the dynamics of the snow on the profiles.

Towards the end of May the watercontent of the snow on the platform in glacier cirque reached in 1959 571 mm, and on the lower (glacier's tongue)-340 mm. During the entire month of June the watercontent of the snow on the tongue was abruptly falling, whereas in the glacier's cirque this fall was noticed only in the last decade of the month.

The general course of snow accumulation during the winter of 1958-1959 on the platforms, situated on the side moraines of the glaciers on the western and eastern side of the valley, fully coincides with the analogous occurrence on the platform established on the tongue of the glacier Zentralny.

The largest thickness of the snow on the left (western) side moraine of the glacier Zentralny, at the beginning of the third decade of April amounted to 151 cm, the watercontent—317 mm, on the right (eastern) side moraine of the glacier Igly Tuyuksu, respectively, 114 cm and 285 mm, while on the glacier — 144 cm and 317 mm. The western platform, in comparison with the eastern is situated in more favourable accumulation conditions, because on this platform is being accumulated snow, that is swept over a small spur (Kypregelsky Gryvka) from the glacier Molodeshny.

Owing to the bedding surface the snow on the western and eastern platforms, in comparison with the glaciers, is removed approximately 10-12 days earlier. The glacier renders a cooling influence upon the snow cover, assisting in its preservation.

On the frontal moraine the snow is usually deposited in hollows, in chutes between the moraine ridges, on the shaded slopes of hills, and on the bottom of lakes, where it is swept from higher and convex-shaped moraine regions which are free from snow during the entire winter.

The largest average thickness of the snow cover on the route A, that was laid across the valley, reached 57 cm at the beginning of May 1959 with a water content of 154 mm, and at the same time the maximum thickness equaled to 115 cm, the minimum, 14 cm.

On the route B that crosses the frontal moraine of the glacier Molodeshny, the maximum of the average heights of the snow cover was observed also at the beginning of May. It equal to 68 cm with a water content of 197 mm, and on the tongue of the glacier Molodeshny, respectively, 111 cm and 278 mm.

The smaller quantity of snow on the moraines is accounted for by the fact that owing to their stony surface, the snow cover, that usually settles here at the beginning of winter later than on the glaciers is exposed earlier to melting and evaporation.

The absolute maximum of the thickness of the snow on the route B reached 135 cm, the minimum-14 cm.

From consideration of the dynamics of the snow cover on slopes with northern and southern exposures, greater thickness and watercontent of the snow is observed to prevail on the northern platform.

The accretion of the snow on the north platform was noticed 10 days earlier than on the south platform.

Of a special interest are the data about the average thickness, density and watercontent of the snowy cover at the end of May-the beginning of June when all of them are characterized by largest values. (Table 5. Fig. 2).

TABLE 5

Average thicknesses, densities and water contents on different profiles of the Tuyuksuysky glaciers on the First of June 1958 and 1959.

		Profiles									
		Gl. Zentralny								Gl. Igly Tuyuksu	Gl. Molodshny
	Years	1	2	3	4	5	6	7	8	1	2
Thickness of snowy cover (cm)	1958 1959	184 77	173 64	182 75	198 109	200 125	193 134	203 149	137	186 —	281 —
Density of snow gr/cm ³	1958 1959	0,37 0,30	0,37 0,30	0,34 0,31	0,36 0,34	0,35 0,31	0,34 0,29	0,32 0,31	— 0,30	0,30 —	0,30 —
water contents (m/m)	1958 1959	681 231	640 207	619 232	716 371	700 388	656 389	650 462	— 411	558 —	843 —
											633 252

Notice : On the glacier Ygly Tuyuksu large thickness and watercontent on the second profile are accounted for by the presence of snow-drifts.

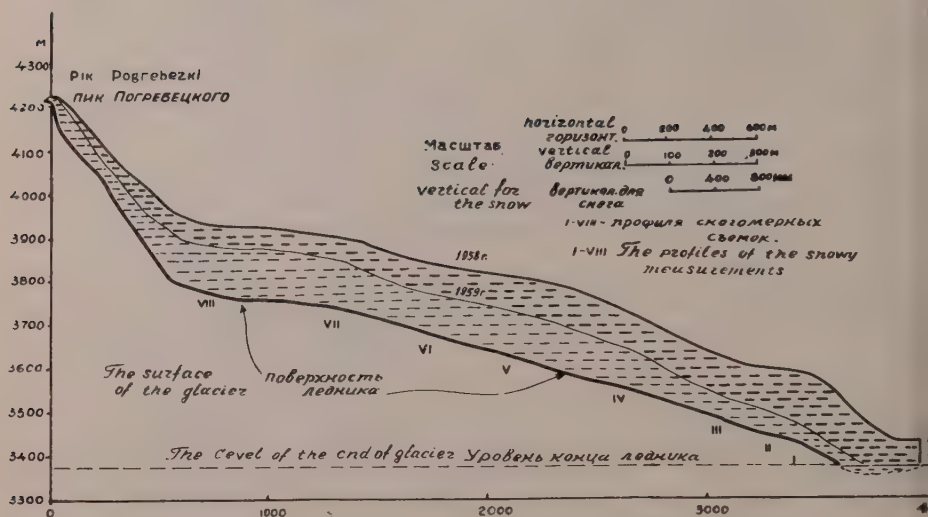


Fig. 2 — The longitudinal section of the snow cover in the water layer, (mm) along the axis of the glacier Zentralny on the First of June 1958 and 1959.

On the crest of the peak Pogrebetsky, the thickness of the snow cover usually reaches 100 cm at the end of the winter, with a density of 0,40-0,45; as a rule, the snow here is swept off into cirque of the glacier.

Utilizing the data of snow-measuring surveys that been carried out on all profiles on the glacier Zentralny, under consideration of the snow-accumulation on separate platforms, an attempt was made to draw a map of the distribution of the water on June 1, 1959 on the entire glacier surface, except the difficulty accessible and dangerous parts, where observation of the snow had not been made.

The isolines of the water content in the snow cover that are drawn on the map are stretched principally along the glacier axis in the direction of the prevailing winds, which are subject to the influence of orography.

On the map referred to, attention is drawn to two particularly pronounced regions of snow accumulation in the glacier cirque and in the middle part of the tongue. An explanation of such distribution of the snow was given by us above.

The isolines of the least water content characterize the regions with convex relief from which the snow is blown, and also regions severely cut by wide and close cracks, into which the snow is swept from the dividing ice blocks (the right side of the fifth profile).

By planimetry of the surfaces with different water content it is possible to determine the volume of water that is contained in the snow cover at the beginning of the warm period.

The dependence of water content in the snow cover upon the absolute altitude at the end of winter 1958-1959 on the glacier Zentralny is shown in Fig. 4. If in the upper parts of the glaciers during the month of May there takes place an increase of the water content at the expense of snow-falls, so in the lower parts the water content of the snow cover remains invariable owing to the intensive melting of the newly fallen snow and the flowing down of the melt water to the end of the tongues.

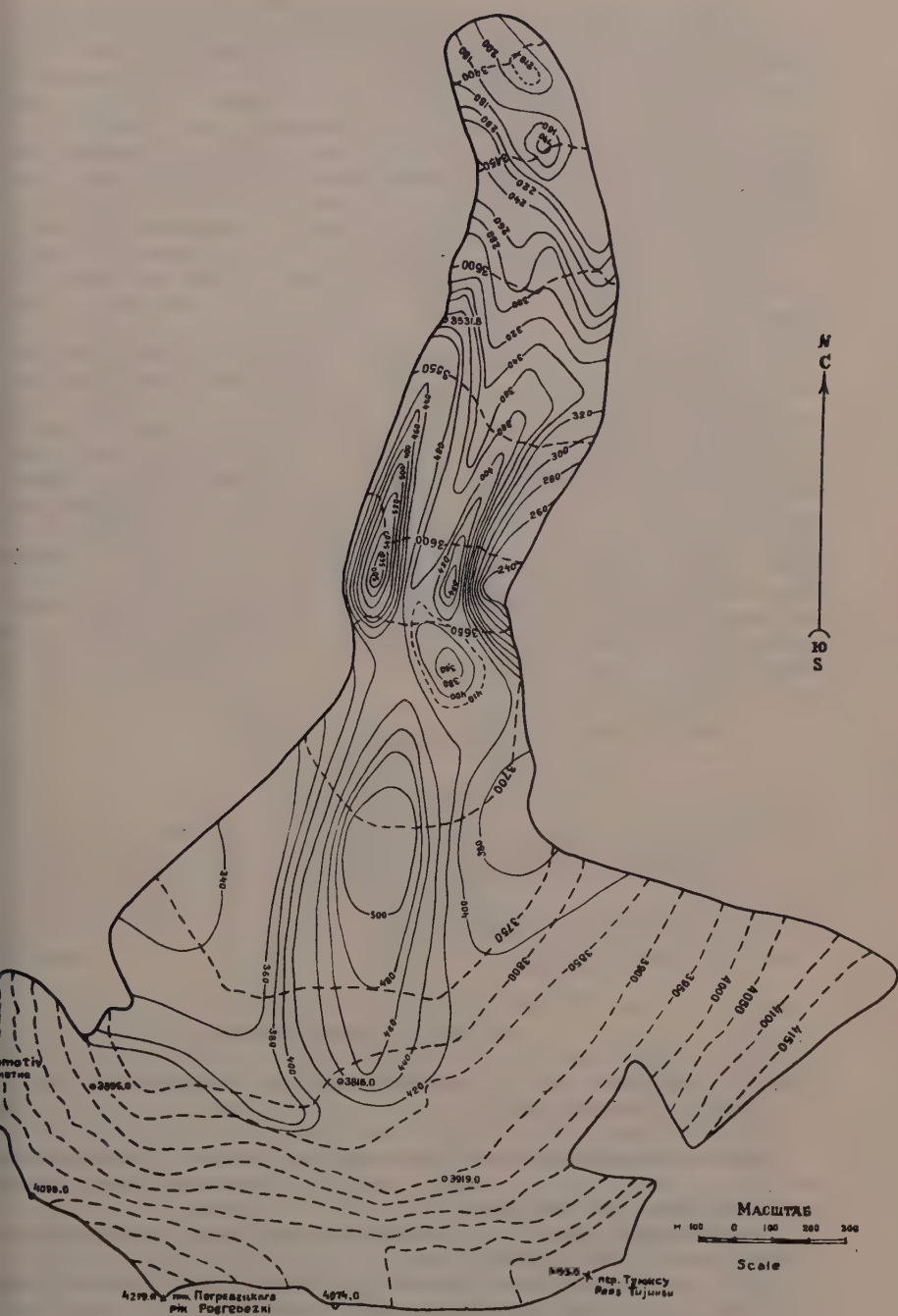


Fig. 3 — Isolines of water content in the snow cover (mm) on the glacier Zentrallyy on the First of June, 1958.

It is necessary to point out that a similar dependence can occur only in those winters when the quantity of precipitation in the cold period on the upper regions surpasses those on the lower regions. If the quantity of precipitation on these regions is equal or nearly so, or when these is even more below than above, the dependence of the water equivalent contained in the snow on the glacier is not subject to the influence of the absolute thickness but is entirely dependent upon the distribution of precipitation in the upper part of the basin, and upon the action of the winds. Such a condition in particular, has been registered in 1958. The quantity of snow, accumulated in the winter of 1957-58 considerably surpassed that of 1958-59. On June 1st 1958 the average water content that has been weighed along the glacier, amounted to 672 mm, but on the same date of 1959 is 383 mm, i.e. 43% less. According to preliminary calculations, on the glacier Zentralny, on June 1st 1958 was concentrated about 2,1 mln m³ of snow, converted into water, but on the same date of 1959, 1,2 mln m³.

As to temperature conditions, the winter of 1958-59 on the glacier during the period of October-May was in comparison with the preceding winter, colder by an average of 0,8°.

The summer period (June-September) of 1959 was by 0,8° warmer than that of 1958. The warmer winter of 1957-58 was also more snowy. The thick snow cover and cool summer of 1958 had played a positive part in the life of the Tuyuksuysky glaciers. The snow line on the glacier Zentralny did not retreat this year higher than 3,660 m over the sea level. In 1957 the snow line on the mentioned glacier reached a level of 3,760 m, this being the result of a light snow-fall winter of 1956-57; although the summer months of 1957 were a great deal cooler than the same months of 1958, and especially those of 1959. The Summer of 1959 was distinguished by high temperatures, which in conjunction with the small quantity of snow accumulated on the glaciers during the preceding winter, have led to an elevation of the snow line to a level of 3,875 m.

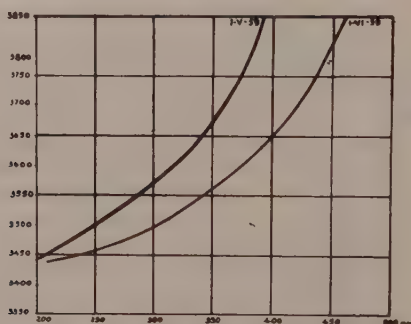


Fig. 4 — The dependence of water content in the snow cover upon the absolute altitude of the glacier Zentralny.

By the said extreme positions of the snow line on the glacier, the ablation surface reached in 1958 only 0,77 Km², and in 1959 2,074 Km², compared to a general surface of the glacier of 3,03 Km².

The fluctuation of altitudes of the snow line on the glacier Zentralny noticeably exerts an influence upon its mass balance.

By its lowest position in 1958, according to preliminary calculations, the glacier lost during the ablation period something over 0,41 mln m³ of ice, converted into water. By the end of the melting period, on the accumulation area there remained accumu-

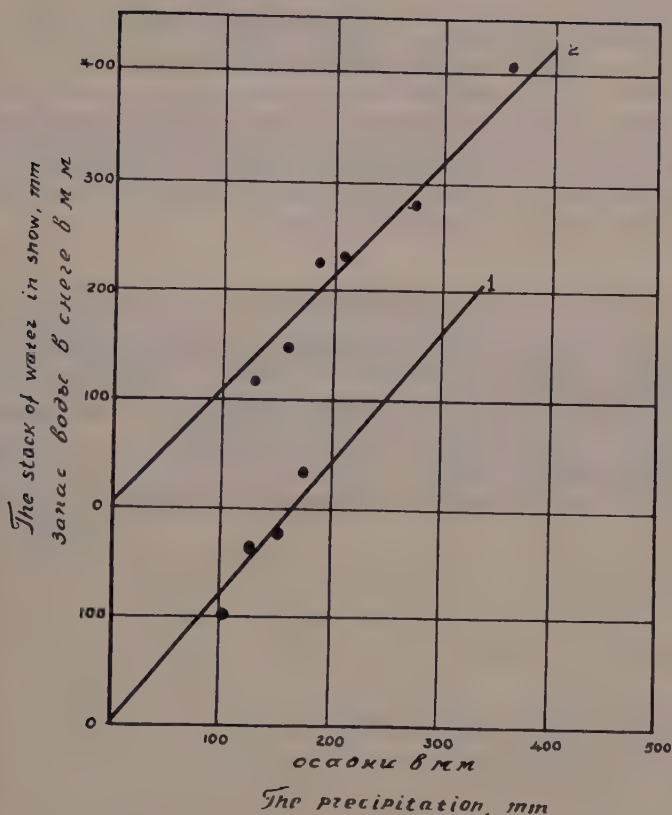


Fig. 5 — The dependence of water content in the snow cover upon the corresponding amount of winter precipitation on the tongue (1) and in the cirque (2) of the glacier Zentralny during the period from September 1958 to May 1959.

ated over 0,7 mln m^3 , or 170% of the mass of ice that was lost by melting, which gives an average value of a water layer of 0,31 m that has been added by the area of accumulation.

A large quantity of snow that had fallen on the glacier during the cold period of 1957-58, led in the yearly period to a positive balance.

In 1959, by an elevated position of the snow line the glacier spent for melting about 2,2 mln m^3 of ice, converted into water, and received on a small accumulation area out of the snow cover of the winter of 1958-59, about 0,4 mln m^3 , or 18% of the substance that has been lost during the ablation. The accumulated snow, that remained for refilling the glaciers after the melting period, represented an average area layer of water of 0,42 m.

On the average during 1958 and 1959 the glacier refilled the water equivalent reservoir to an extent of only 42% of the quantity that it originally spent by melting.

Thus, the climatic conditions in the glacier zone of the Zailiysky Alatau are at present unfavourable for glaciation. The retreat of the glaciers, the reduction of their thickness and volume of the water reservoir in a solid form that were noticed before, are being continued at present (1;3).

The substantial influence of the winds upon the distribution of snow on the glaciers was repeatedly pointed out in the present report. As the observations had shown, the stores of snow on the glacier Zentralny surpass the amount of precipitation, registered by the gauges that were set near the snow measuring platforms. The comparison of the maximum amount of water content in the snow cover with the amount of fallen precipitations from the moment of its setting is given in fig. 5. About 12-15% of the snow accumulated on the glacier during the winter can be attributed to snow that was brought in from the surrounding slopes and swept over from the upper parts of the neighbouring basins.

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THE PRELIMINARY DATA ON THE FORMATION OF ICE IN THE ZONE OF ACCUMULATION OF THE TUYUKSU GLACIERS

K.G. MAKAREWICH and G.A. TOKMAGAMBETOV (U.S.S.R.)

Glaciological group of the Geographical Department of the Academy of Sciences of the Kazakh S.S.O.

SUMMARY

The observation of the formation of ice in the zone of accumulation of the Central Tuyuksu glacier in the Zailiysky Alatau (the North Tien Shan) began in autumn of 1956 in the period of the preparation for the investigations on the program of IGY. Ten wooden stakes of 3 or 4 m high were sited along the foot of the slope in the cirque of the glacier. The surface of snow after the period of ablation the next winter was marked yearly by the slim layer of the wooden sawdust. The distance between the extreme stakes in the western and eastern sides of the cirque is almost 600 m.

The observations of the accumulate thicknesses of snow by the excavation of them on the mark parts were carried out during next three years. The volume weight, porosity of snow, firn and ice on the different layers, their structure and texture, heat physical and mechanical properties were studied with the help of specimens from the excavations.

One excavation and a hole were set up for the same aims at a distance of some hundred meters from the afore-mentioned stakes. The meteorological conditions, solar radiation, and temperature regime of ice along the depth of 22 m in the zone of accumulation were studied at the same time.

The first results of the observations may be summed up as follows :

1) the zone of accumulation of the Central Tuyuksu glacier stretches approximately on 1/3 length and is equal to 1/2 of the whole area.

2) the infiltrate-congelate type of the formation of ice take place in this zone.

3) the cycle of the formation of ice lasts two years from the moment of the beginning of accumulation of the season snow. Within the first year the season snow transformates into firn, the processes of firnification take place within the next year. At the end of this year the firn transformates into ice. In some cases this process is completed at the beginning of the third year (illustrated by snow-firn profiles). The maximum firn layer at the foot of the slope in the cirque of glacier is equal to 160 cm, the minimum — 14 cm.

4) Three kinds of ice are observed in the zone of accumulation : transparent infiltrate, untransparent infiltrate and infiltrate congelate ice. This is confirmed by the structure investigations (stereogrammes).

The report is accompanied by the tables of the volume weight, porosity, heat conductivity, and hardness of snow, firn and ice for the different parts of the zone of accumulation.

RÉSUMÉ

Les observations sur la formation de la glace dans la zone d'accumulation du glacier Toujouksou dans l'Ala-Tau transilique (Tien Chan septentrional) furent commencées en automne 1956 au cours de la préparation des travaux de recherche conformément au programme de l'A.G.I. Dix jalons en bois de 3-4 mètres de haut furent plantés le long du pied du mur occidental du cirque du glacier. Autour de chaque jalon, dans un cercle d'au moins 3 mètres de diamètre on marqua chaque année, par une mince couche de sciure de bois, la surface de neige restée après la période de la fonte jusqu'à l'année suivante. La distance entre les jalons extrêmes occidental et oriental se chiffrait à près de 600 mètres.

Durant les trois années suivantes des observations ont été faites sur l'épaisseur d'accumulation de la neige à l'aide de fouilles de recherche creusées sur l'emplacement des aires marquées. Le volume spécifique, la porosité de la neige, du névé et de la glace sur différents horizons, leur structure et texture, leurs propriétés thermophysiques et mécaniques ont été étudiés sur des échantillons tirés de ces fouilles. A ces mêmes fins, une fouille et un sondage furent exécutés à une distance de quelque cents mètres

des jalons mentionnés plus haut. On étudia simultanément dans la zone d'accumulation les conditions météorologiques et de radiation ainsi que le régime de température de la glace jusqu'à une profondeur de 22 mètres.

Les premiers résultats des observations se ramenèrent aux positions essentielles suivantes :

1. La zone d'accumulation s'étend approximativement sur un tiers de la longueur du glacier central de Toujousou et se chiffre à près de la moitié de sa surface totale.

2. Dans la zone mentionnée on se trouve en présence d'une formation de glace du type infiltration-congélation.

3. Le cycle de formation de la glace se déroule durant deux ans à partir du début de l'accumulation de la neige saisonnière. Au cours de la première année, la neige se transforme en névé, l'année suivante se déroule le processus de métamorphose de la couche de névé qui à la fin de l'année se transforme en glace. Dans certains cas cette transformation se termine au début de la troisième année (coupes transversales de la couche de neige et de névé). L'épaisseur maxima de la couche de névé au pied du mur occidental du cirque du glacier atteint 160 cm, alors que l'épaisseur minima est de 14 cm.

4. On rencontre trois sortes de glace dans la zone d'accumulation : glace transparente d'infiltration, glace opaque d'infiltration et glace d'infiltration-congélation, ce qui est confirmé par les recherches structurelles (stéréogrammes).

Le rapport est accompagné de tableaux, où l'on trouve des données sur le volume spécifique, la porosité, la thermoconductibilité et la résistance de la neige, du névé et de la glace dans différents secteurs de la zone d'accumulation.

The Tuyuksuysky glaciers are situated on the northern slope of the ridge of the Zailiysky Alatau (Tien Shan), 35 Km away from the capital of Kazakhstan, Alma-Ata. The highest altitudes in the glacier region reach 4,410 m. The main glacier Zentralny Tuyuksuysky has a length of 5,1 Km (including the frontal moraines) and an area of 3 Km².

The climatic conditions here are typical for the high-mountain glacier zone. The average year's temperature is between $-4,2^{\circ}$ (3,420) to $-6,5^{\circ}$ (3,750 m). A more detailed characterization of the meteorological elements is given in the report dealing with the distribution of the snow cover on the mentioned glaciers, which is included in the present edition. There is available a plan of the Tuyuksuysky glaciers and also a map of the glacier Zentralny. In this report that give a general idea of the glaciation of the observed region.

In 1956, at the foot of the back cirque wall of the glacier Zentralny Tuyuksuysky there were installed 10 brightly painted wooden stakes of a length of three and four

TABLE 1

Absolute altitudes and distance between accumulation stakes (Profile 8)

Number of the stakes	200	201	202	203	204	205	206	207	208	209
Absolute altitude (M)	—	—	3788	—	3778	—	3778	3780	3781	3785
Distance between stakes		40	42	60	64	65	60	60	38	126

meters. Their lower end was bored into the firn, the upper surface of which, before the snow accumulation had begun, was marked by a thin layer of wooden sawdust of a diameter of 3 meters. The marking regions were located in such a way that the snow avalanches of the slopes of the peak Pogrebezky should not reach the stakes. Owing to this their prolonged safety was secured. Besides that, there were taken into consideration the fluctuations of the thickness of the snowline on the glacier with such a margin, that even at its highest position it should be possible to obtain data of the extent of the snow accumulation on the glacier at the end of the ablation period.

During 1957-59, at the end of each ablation period the surface of the seasonal snow that remained in the next winter was newly marked by wooden sawdust.

As supporting data for the studying of the accumulation zone, studies of the traverse profiles, including levelling, determination of motion speed, extent of melting snow-measuring surveys, etc., were made. Information of accumulation stakes and transverse profiles are given in the tables 1 and 2.

TABLE 2

Some data of transverse profiles at the upper part of the glacier Zentralny Tuyuksuysky

Profile	Number of supporting points	Average altitude of profile (M)	Average distance between profiles and watershed	Width of profile (M)	Average inclination angle (degree)	Remarks
8	10	3782	700	585	6°	The inclination angles are given for region in direct vicinity of stakes
7	12	3776	1300	700	4°	
6	8	3675	1710	525	7°	

In September 1957, in the region of the accumulation stakes, pit-hole no. 4 was excavated for observation of the firn and ice temperature regime to a depth of 25 meters (absolute altitude about 3,800 m). For the same purpose, at a distance of 500 m from the accumulation stakes, down the glacier, almost in the centre of the glacier cirque, in August of the same year, another pit-hole no. 3 was excavated at a depth of 23 meters (absolute altitude 3,750 m). Beginning on the first of July 1957, regular meteorological observation and solar radiation measurements have been made in the glacier cirque. At the same time observations of snow fall, dynamics and distribution of the season snow cover in the glacier cirque, and also its thawing and participating in the refilling of the glacier mass were begun.

Before proceeding to describe the accumulation zone of the glacier Zentralny it was necessary to determine its extent. As one of the most widely adapted criteria for division of glaciers into ablation and accumulation zones by area, by altitudinal intervals, or by length of the glacier, there is the visible position of the snow line, or, as it is often called, the firn line. But such a division does not correspond to the real

correlation of the mentioned zones because the processes of the ice formation on the glacier take place at a distance from the snow line and the glacier accumulation region surpasses considerably the accumulation zone that is being singled out according to the above mentioned criteria.

According to the observation data on the glacier Tuyuksu, the firn line can be considered the edge between the snow covered and infiltrated ice with a smooth surface on one side, and the ice of deeper layers on the other side, the surface of which had been heavily transformed by melting. Proceeding from this determination, the accumulation zone of the glacier Tuyuksu reaches at the most one third of its length and equals approximately to one half of its entire area. Therefore, in the matter of firn line, the authors fully adhere to the point of view expressed in the works of P. Shumsky and G. Avsuk (1955, 1956).

In the singled accumulation zone observations were made of hardness, density, porosity and thermophysical properties of firn and ice in the surface layers.

These observations were made along the profiles VI, VII, VIII and owing to this, any observation point was topographically mapped. Moreover, studying the physical properties of ice by analogous observations on the low parts of the Tuyuksu glaciers allows utilization of the obtained data for comparison with the characteristics of the accumulation zone. The data are shown below in the table 3.

The significances of average values of hardness of firn and ice in the accumulation zone quoted in the table 3 show that the firn possesses a hardness from 0,113 to 0,138 Kg/cm², the infiltrated ice—from 0,436 to 784 Kg/cm², but the depth-ice that is emerging on the surface near the firn line—from 1,197 to 3,175 Kg/cm². The coefficients of temperature conduction and thermoconduction vary accordingly.

The changes of the density of the ice from 0,908 to 0,894 gr/cm³ lead to the change of the mentioned coefficients, the former from 0,99 to 0,89. 10⁻² cm²/sec., the latter from 4,5 to 4,0. 10⁻³ cal/cm, degree-sec. For the firn, by a density from 0,80 to 0,54 gr/cm³ the temperature conduction varies from 0,87 to 0,72. 10⁻² cm/sec., but the thermoconduction from 2,5 to 2,0. 10⁻³ cal/cm degree—sec. Thus, in the accumulation region the processes of ice formation proceed over a considerable area beyond the edge of the snowline, reaching an absolute altitude of 3700-3720 m, and a distance from the watershed of 1400-1540 m.

Besides the routine studies, in the pit-hole no. 3, specimens were taken of firn and ice from different horizons of the accumulation zone at a depth up to 22 m. With the aid of these specimens it was possible to carry out chrySTALLINE—optical studies, determination of density and porosity. In the same region a 5 m deep pit was opened, where, besides the above mentioned determinations, experiments on mechanical and thermophysical properties of the ice were carried out.

The cross section of the ice shown in fig. 1 characterizes as follows: the ice layers are placed horizontally and parallel to each other. Also, below the surface separate firn layers were found that reveal the presence of firn lenses at different depths. It is necessary to point out that the latter were not noticed in the pit. These lenses were once isolated from the influx of the melt water from above, by waterproof layers of ice. In the depth of the first 10 meters from the surface several such firn layers were counted, but beneath them their number diminishes noticeably, whereby a regular augmentation of the firn density and thickness of the ice layers and a diminishment of the thickness of the firn layers were noticed. In the upper horizons the density amounts to 0,57, in the lower—0,73 gr/cm³.

The fact of the presence of firn at a depth of 20-21 m shows that its transformation into ice has not been fully terminated and the processes of ice formation under the conditions of a complete isolation from external factors are being completed at a considerable depth by way of subsidence and paratectonic recrystallization.

No. of the transverse profiles	No. of stakes	Characterization of specimen	Density cm ³ /sec	Conduction of temperature cal/cm	Thermo conduction gr/cm ³ degree sec.	Hardness of ice and firm			Remarks
						Maxim.	Minim.	Average	
1	2	3	4	5	6	7	8	9	10
VI	302	ice	0,888	0,90.10 ⁻²	3,9.10 ⁻³	9,840	1,640	3,175	For conducting the route observations for determination of hardness.
	303	ice	0,887						
	304	ice	0,896	0,94.10 ⁻²	4,2.10 ⁻³	4,920	1,640	3,056	
	305	ice	0,910	0,98.10 ⁻²	4,5.10 ⁻³	3,780	0,984	1,929	
	307	ice	0,920	0,94.10 ⁻²	4,3.10 ⁻³	1,640	0,820	1,197	
	308	ice	0,894	0,89.10 ⁻²	4,0.10 ⁻³	2,460	1,406	1,776	
VII	1	ice	—			0,820	0,518	0,586	a load equalizing to 2 kg was used.
	2	ice	0,887			0,878	0,665	0,762	
	3	ice	0,902			0,517	0,428	0,468	The diameter of the ball: for ice 2,33 cm, for firm 3,97 cm.
	4	ice	0,897			0,593	0,434	0,494	
	5	ice	0,872			0,529	0,400	0,462	
	6	ice	0,891			0,566	0,394	0,444	
	7	ice	0,893			0,497	0,407	0,436	
	8	ice	0,892			0,757	0,523	0,601	
	9	ice	0,899			0,769	0,647	0,709	
	10	ice	0,884			0,820	0,656	0,226	
VIII	201	firm	0,70	0,67.10 ⁻²	2,5.10 ⁻³	0,138	0,095	0,113	The number of determinations in each point amounted to 10
	204	firm	0,56	0,72.10 ⁻²	2,0.10 ⁻³	0,156	0,115	0,134	
	206	firm	0,80	0,87.10 ⁻²	3,5.10 ⁻³	0,167	0,124	0,147	
	207	firm	0,54	0,72.10 ⁻²	2,0.10 ⁻³	0,170	0,105	0,127	
	208	firm	0,70	0,66.10 ⁻²	2,3.10 ⁻³	0,142	0,109	0,111	
	209	firm	0,56	0,75.10 ⁻²	2,1.10 ⁻³	0,156	0,110	0,132	
	210	firm	0,70	0,66.10 ⁻²	2,3.10 ⁻³	0,164	0,113	0,138	

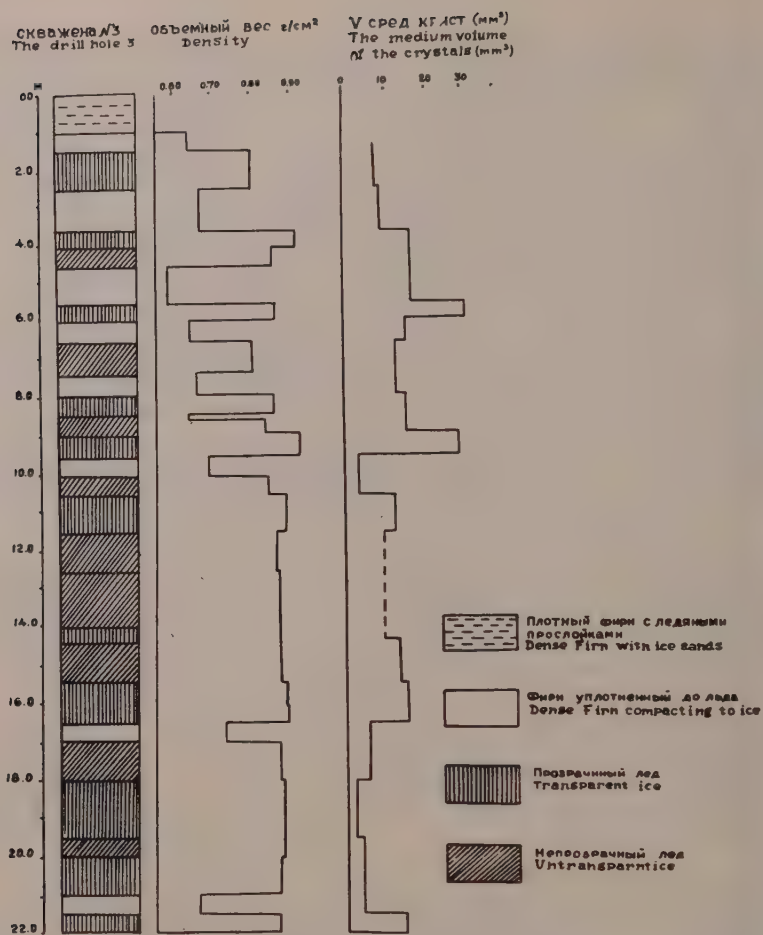


Fig. 1 — A section of ice in the cirque of the Tuyuksu Glacier (no. 4 drill hole)

According to data of electrometric studies in the region of the pit-hole court-
teously left at our disposal by B. Borovinsky, there was noticed a zone of alternating
apparent resistance that may be explained by alternation of the ice and firn layers.
By these observations the zone of pure ice was traced to a depth of 21-23 meters,
where the said resistances are characterized by constant values. According to the
textural description of the specimen, two kinds of firn were observed: a fine-grained
firn with grains of 0,2-0,5 mm, a density of 0,64-0,70 gr/cm³, and a porosity of 0,25-
0,30%, and a coarse-grained firn with grains of 1,5-2,0 mm, a density of 0,57-0,65 gr/cm³
and a porosity of 30,0-0,38%. The ice here was also of two kinds: transparent infil-
tration ice that is widespread throughout the entire cross section, with a density of
0,85-0,91 gr/cm³ and a porosity of 0,01-0,08‰, and a dull, untransparent infiltration
ice that is characteristic for the upper horizons, with a density of 0,84 to 0,88 gr/cm³
and a porosity of 0,4-0,9‰.

In outline the structure of the infiltration ice is orientated allotriomorph-grained
with comparatively small-crystals. In the upper horizons the predominant orientation

of axes in relation to layers is clearly expressed (Fig. 2). With depth, there was observed a large diffusion of the main axes as far as the parallel plane of the layers, which, as P. Shumsky (1955) remarks, « have served as crystallization centres of the firn grains».

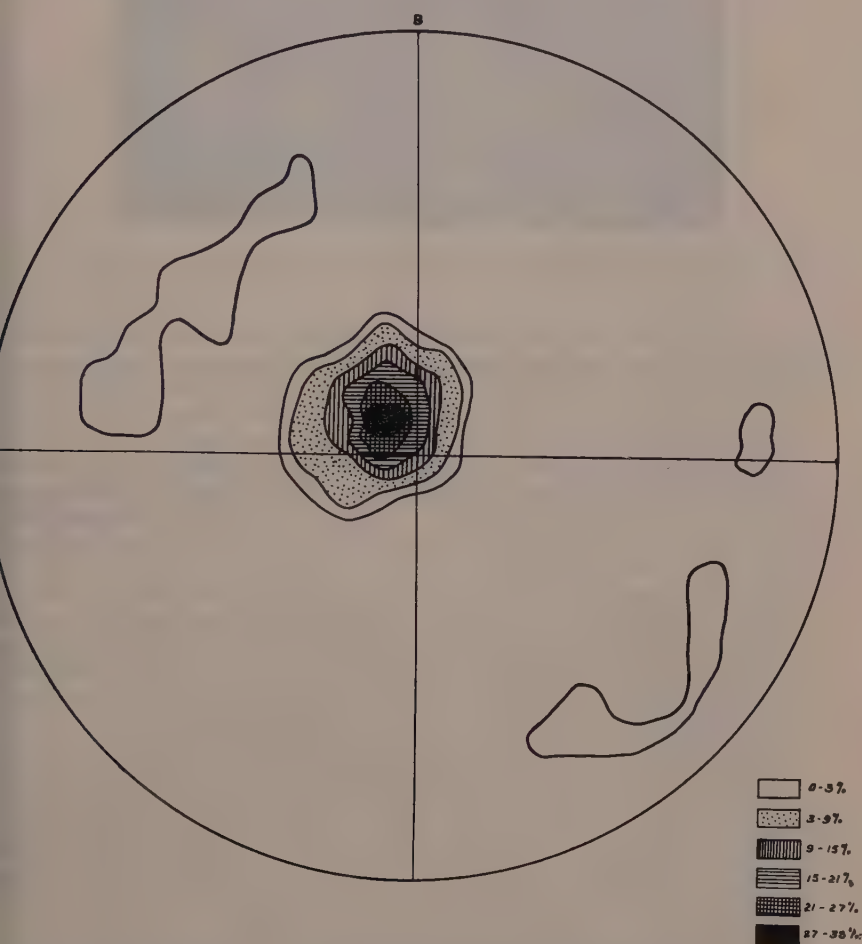


Fig. 2 — Stereogramme of the crystallographical orientation of the infiltration ice.

The crystals are mainly isometrical with distinctly expressed facets, especially on small crystallized structures.

The predominant form of the crystals is the Pentahedron (Fig. 3). Bubbles of rounded form are between the crystals on the facets. The ice crystals up to a depth of 5 meters have a tendency to increase their volume from 8,4 to 31,3 mm³. There is noticed a relative constancy of the dimensions, at the depth of 5 - 8,5 m. At the ninth meter a sharp increase of the crystal volumes up to 35 mm³ follows at the tenth

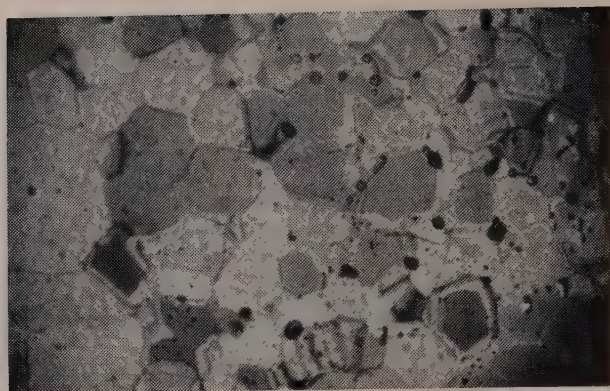


Fig. 3 — The allotriomorphic-grained structure of the infiltration ice + 4.

meter—a sharp sinking to $2,7 \text{ mm}^3$ and after that a relative constancy is again being observed. Such leaps are conditioned, apparently, by the formation peculiarity of the infiltration ice.

The processes of ice forming in accumulation zone of the Zailiysky Alatau glaciers follow the principal regularities of the infiltrate-congelation type.

The temperature conditions are here characterized by the following figures (according to the observation results from September 1958 to August 1959): the average yearly air temperatures fluctuate within -6° ; the temperature of the coldest month (January or February) is contained in the section $-1,3, 8-15,9^\circ$, and of the most warm month $+2,8-+3,3^\circ$ (July-September).

The temperature regime of the ice is closely bound with air temperature and according to the distribution of the temperatures at depth in ice three principal zones are distinguished: the surface, the middle zone, and the deepest zone. The first one spreads from the surface down to 10 meters, the second one—from 10 to 20 meters, the third one—below 20 meters. The stratification of the temperatures in these zones fully coincides with the description of such by G. A. Avsiuk (1956).

The entire depth of the ice up to 23 meters has a negative temperature.

In the surface horizon the course of the ice temperatures strictly follows the course of the air temperature with one month's delay. During one year's accumulation cycle of firn precipitation, from September 1958 to August 1959, the coldest month February had a temperature of the air of $-15,9^\circ$, and the warmest month, July $+2,8^\circ$, whereas, the temperature of ice at the depth up to 1 meter from the surface observed in March was $-7,0-6,7^\circ$, in August, $-1,6-2,1^\circ$. The values of temperature up to the depth of 23 m for separate seasons is given in table 4.

Owing to transmission of the cold from the inner layers to the surface of the glacier and also in view of the presence of a large quantity of snow and comparatively low temperatures in the summer time, in the region of the pit-hole no 3, at the end of the warm period of 1958, a layer of infiltration ice of a thickness up to 50 cm has been formed.

In the table attention is drawn by the abrupt change of temperatures at the depth of 20-21 meters which is most probably the result of the presence of the firn band.

At the foot of the back cirque wall of the glacier Zentralny, in the direct vicinity of the accumulation stakes, in the autumn of 1958 and summer of 1959 were opened several pits up to 5 m deep. In the process of this work the limits between the remnants of the year's accumulation, that were marked by sawdust, distinguished very clearly.

TABLE 4

Depths in Months	0.0	0.1	0.2	0.3	0.5	0.75	1.0	1.5	2.0	2.5
October 1958	- 2.7	- 2.6	- 2.7	- 2.75	- 2.7	- 2.3	- 3.55	- 2.7	- 2.75	- 2.4
March 1959	- 7.0	- 7.0	- 7.0	- 7.1	- 7.2	- 7.2	- 6.7	- 6.4	- 5.9	- 5.5
May 1959	- 4.5	- 4.6	- 4.5	- 4.6	- 4.6	- 4.6	- 5.0	- 5.2	- 5.2	- 5.2
August 1959	- 1.6	- 1.8	- 1.5	- 1.56	- 1.4	- 1.7	- 2.1	- 2.6	- 2.9	- 3.2

Depths in Months	3.0	4.0	6.0	8.0	10.0	15.0	20.0	21.0	23.0
October 1958	- 2.6	- 3.1	- 3.3	- 3.45	- 3.10	- 2.7	- 2.3	- 1.4	- 1.5
March 1959	- 5.1	- 4.5	- 3.4	- 3.0	- 2.9	- 3.25	- 2.25	- 1.3	- 1.5
May 1959	- 5.0	- 4.6	- 3.6	- 3.3	- 2.9	- 3.1	- 2.3	- 1.4	- 1.5
August 1959	- 3.3	- 3.35	- 3.4	- 3.45	- 3.2	- 3.1	- 2.4	- 1.5	- 1.6

The cross section of the principal pits, situated on the west, central and east side of the cirque are shown in the Fig. 4.

In the cross section of the pit no. 1 which has been laid in the western part, there was noticed that beneath the snowy-firn are bedded layers, slightly inclined northwards, that are represented by transparent and untransparent infiltration ice

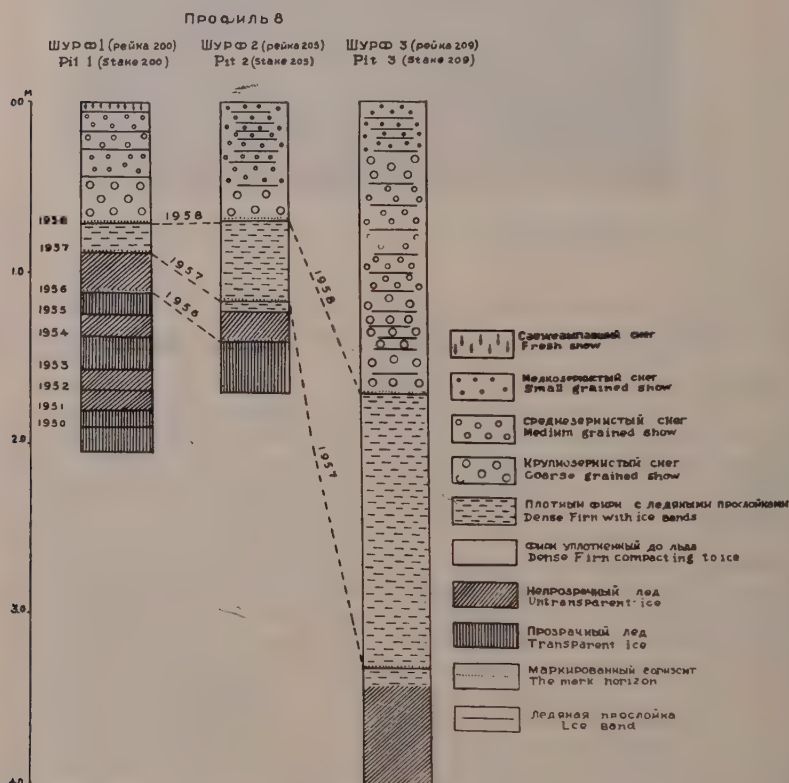


Fig. 4 — Firn-ice profiles in the pits dug next to stakes of accumulation of the Tuyuksu Glacier.

The ice layers, bedded below the horizon of 1956 are shown conditionally.

Above the 1958 horizon there is a layer of small-grained and coarse-grained snow which is divided by icy bands of a thickness of 72 cm that were formed as a result of radiation melting. Between the 1958 and 1957 horizons there is a layer of a 16 cm thickness, but between 1957 and 1956 there is an untransparent infiltration ice layer of a thickness of 24 cm. In the central part of the accumulation area, snow in 1959 was determined to be of a thickness of 70 cm, but the firn layer deposited beneath the 1958 horizon, that characterizes the rest of the yearly accumulation of 1957-58, equals 45 cm. Beneath the 1957 horizon there was preserved a firn layer of 2-3 cm, that towards the end of the ablation period converted into ice of the 1956-1957 accumulation of a general thickness of 24 cm. On the east side of the cirque the greatest accumulation depths, are noticed, due to transference of the snow from the Tuyuksu pass region, which is deposited at the foot of the slope to depths of several meters.

ready in 1959 the accumulation stakes were completely covered with snow. The pit opened in this place is characterized by the following figures. The accumulation of 1958-59 comprised, towards the end of July 170 cm of snow of different texture and with numerous ice streaks. The layer of a year's accumulation during 1957-58 reached 100 cm. There is deposited a layer of ice below it. For identifying the texture horizons the authors made use of the recommendation of P. Shumsky (1955). The firn was considered grained snow, irrespective of the age in which the idiomorphic structure that is natural for snow had been lost, but for the ice the connected pres which are natural to the firn were also lost. By observations during the I. G. Y. some peculiarities of the accumulation regime, that are first of all dependent upon the climatic factors, were fixed.

The winters of 1956-57 and 1957-58 were the most snowy, although the former was inferior to the latter. The winter of 1958-59 appeared to be of little snow. As to temperature conditions, the winter of 1957-58 was a warmer one than that of 1958-59, but the warm period of 1958 was cooler than the same period of 1959. All this influenced the altitude of the snow line on the glacier. In 1957 it reached a level of 3,760, in 1958—3,660, in 1959—3,875 m. The fluctuation of the snow line and the peculiarities of the distribution of the snow cover on the glacier effect the processes of the ice forming, the spatial position of accumulation zone and the thicknesses of the yearly accumulation layers.

Thus, on the cross section pits in the accumulation zone, the following picture has been observed. Towards the end of the ablation period of 1959 in the region of pits 1 and 2 in the western and central parts of the section, a complete melting of snow that was deposited in the winter above the 1958, horizon was noted. The melt water, saturating the lower firn layer accelerates the ice forming processes, thus resulting in the rise of the surface of the infiltration ice.

As much as in 1959 the melting period lasted until the 5 October, a part of the newly formed infiltration ice melted, but owing to the penetration of the cold in to the ice, the melt water partially froze on the surface of the glacier in the accumulation region, thus forming snow covered ice with a smooth surface. The lower firn layer that belongs to the accumulation of 1957-58 during the warm period of 1959, was turned into ice. In the eastern part of the glacier, where the extent of snow accumulation was considerably higher during the said period of 1959 a firnification of snow had taken place, but the firn layer of the 1957-58 accumulation, that is situated below, was exposed to the actual processes of ice forming, which may be completed only in the third year, and possibly in the fourth year.

Thus, even under conditions of comparatively low altitudes and in a region of comparatively high air temperatures, the ice forming cycle can be completed during different intervals of time. It is dependent on the snow accumulation amount, all other conditions being equal.

This is confirmed by observations of the temperature regime in pit-hole No. 4 that was opened in the neighborhood of pit No. 3. In the winter of 1957-58 the temperature of the surface horizon, irrespective of the presence of a snow bedding, had sunk to $-18,8^{\circ}$. During the period of 1958 a large amount of snow has been accumulated here, out of which 160 cm remained for the next winter. The temperature of the same surface horizon, on which this snow had been accumulated, rose in the summer of 1958 to $-0,1^{\circ}$, which is, first of all, connected with the infiltration of the melt water. In the next winter of 1958-59 the temperature sank on the same horizon to $-7,9^{\circ}$ only, and at that time the rest of the yearly accumulation of 1957-58 experienced the processes of firnification. A new accumulation of snow from the end of the ablation period of 1958 to the approach of the warm period 1959 increased the thickness of the observed accumulation. The melt water, however, had an opportunity to again penetrate into the above mentioned horizon and to raise its temperature to 0° .

Towards the end of the melting period of 1959 the rest of the yearly accumulation of 1956-57 was definitely transformed into ice.

CONCLUSIONS

1. The accumulation zone of the Tuyuksu glacier extends approximately 1/3 of its length and 1/2 of its entire area.
2. In this zone there takes place an infiltrational-congelation type of ice formation while the duration of the process differs and depends, principally, on the following factors:
 - a) the quantity of solid precipitations, that falls during the year;
 - b) the distribution and extent of accumulation, connected with orographic and wind conditions;
 - c) the altitude of the snow line on the glacier;
 - d) the temperature regime of the glacier zone;
 - e) the absolute altitude of the accumulation zone;
 - f) the morphology of the glacier cirque, etc.;
3. The ice forming cycle in the lower parts of the glacier feeding district completed during one year; in the upper parts in separate years a biennial cycle observed, but in some cases the completion of the ice forming cycle takes place also in the third year.
4. The noted peculiarities of the ice formation can be extended to the glaciers of the entire ridge of the Zailiysky Alatau.

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APPLICATION OF THE SPORE-POLLEN ANALYSIS FOR STUDYING THE STRUCTURE OF GLACIERS ON ELBRUS MOUNTAIN

E.S. TROSHKINA and J.V. MACHOVA
U R S S

SUMMARY

1. In the firn area of glaciers snow-fall happens the whole year round, the deposit of flowering plants' pollen taking place simultaneously. Monthly superficial samples of snow, taken at height of 3600 m of the southern slope of Elbrus, made it possible to determine the seasonal accumulation of snow, according to data of store-pollen analyses obtained by investigations of firn-ice sections up to 12 m deep.

2. According to pollen spectra of the sections having been analysed, there have been distinguished the spring, summer, autumn and winter horizons. The spring horizon is characterized by prevalence of the pollen of birch, alder, hazel-nut, elm, oak-tree, hornbeam and other spring flowering plants. The summer horizons are enriched by pollen of dark-coniferous trees (fir, silver fir). The autumn spectra are distinguished by its unstable, even occasional composition of pollen, which finds its way to a glacier. Winds getting stronger at that period, lift the pollen settled after flowering on the mountain slopes. The winter horizons contain the least quantity of the pollen.

3. As a result of determining season deposits of different snow layers, the accumulation of hard substance on the southern slope of Elbrus from 1954 to 1958 was established.

4. Composition of the pollen and spores is the evidence of prevalence by transfer western winds. It confirms point of G. K. Tushinsky's view about the great value of western snow-storm transfer which play an important role in the nutrition of Elbrus glaciers.

5. According to the data of spore-pollen analyses the banding of glacier tongues, in of cthe plastic flow, is related to the primary striation of the ice.

RÉSUMÉ

1. Dans la zone du névé des glaciers la neige tombe tout le long de l'année, en déposant simultanément le pollen des plantes en fleur. Les tests superficiels mensuels de la neige faits à une altitude de 3600 m sur la face sud de l'Elbrouz permirent de déterminer les saisons d'accumulation de la neige en se basant sur les données de l'analyse cryptogamique et par pollinisation lors des recherches sur des coupes de névé d'une profondeur atteignant 12 m.

2. D'après les spectres cryptogamiques des coupes analysées il a été possible de distinguer les horizons de printemps, d'été, d'automne et d'hiver. Les horizons de printemps sont caractérisés par la domination du pollen du bouleau, de l'aune, de la brème, de l'orme, du chêne, du charme et autres essences d'arbres fleurissant au printemps. Les horizons d'été sont fécondés par le pollen des espèces conifères foncées (sapin, épicéa). Les spectres d'automne se distinguent par le manque d'uniformité de leur composition en pollen, composition pouvant même être purement accidentelle. Le pollen se dépose, en général, sur le glacier sous l'action du vent, qui, plus violent à cette époque, soulève le pollen déposé après le fleurissement sur les versants des montagnes. Les horizons d'hiver contiennent une quantité minima de pollen.

3. Par la détermination des saisons de dépôt de différentes couches de neige il a été possible de calculer l'accumulation des substances dures sur la face sud de l'Elbrouz entre 1954 et 1958.

4. La composition du pollen et des cryptes témoigne que ces derniers sont transportés par les vents dominants de l'Ouest ce qui confirme le point de vue de G. Tushinski sur l'importance considérable des mutations occidentales dues aux tempêtes de neige qui jouent un rôle décisif dans l'alimentation des glaciers de l'Elbrouz.

5. D'après les données de l'analyse cryptogamique et par pollination la présence de couches stratifiées dans les langues de glaciers est liée à la stratification initiale de la glace dans le cas d'écoulement plastique.

Glacier formation on the Elbrus was studied in accordance with the program of glaciological investigations of the International Geophysical Year. The method adopted by us was the spore-and-pollen analysis for determining the seasonal snow and ice stratification, the origin of individual ice belts, as well as the size of solid substance accumulations in the investigated glaciers.

The work was carried on mainly on the southern slope of the Elbrus 3600-4100 m high and also on the glacier Karachaul Northern slope. The first task confronting us was to determine the seasonal snow and ice stratification.

Samples for analysis was taken from two crevasses; one of them located on the Garabashi glacier at a height of 3600 m, the other in the firn basin of the glacier at a height of 4000 m (near the Refuge of Nine).

The first crevasse was longitudinal—1.45 m wide, 45.5 m long and about 7 m deep. The ice layers in it were clearly visible, inclined at an angle of 30° and divided by dirtybrown strips 15-16 cm wide. In this crevasse, under cover of a superficial mud crust, there was a horizon of moist solid snow (of 0.56 g. density) 1.90 m thick. Beneath it there was a dirty-brown layer of ice 16 cm wide. Under it comes a horizon of ice 1.30 m thick, changing into a streak of ice 15 cm deep. Still lower, at a depth of 3.50 to 5.80 m we find a thick layer of blue ice. The next horizon 1.00 m deep is separated from the foregoing by a brown strip 15 cm thick and the crevasse ends at a depth of 7.05 m with a brownish horizon 15 cm wide. Further down the crevasse becomes wedged.

The second crevasse, from which samples were taken, is lateral, over 20 m deep and up to 2 m wide. The layers were clearly defined in it, which made it interesting to compare the two section cuts.

The pollen and spores found in the ice and snow-firn strata are brought by the wind. During the period of plant florescence, the sedimentation of their pollen on the surface of glaciers takes place. The sequence of various tree florescence, the pollen of which is found in the analysed samples may be seen in the table (see Table 1)

TABLE N 1

Blooming Period of Various Types of Trees (Data supplied by A. A. Grossheim)

Blooming Period	Kinds of Trees
February-March March	<i>Alnus barbata</i> <i>Corylus avellana</i>
March-April	<i>Corylus iberica</i> , <i>Betula Raddeana</i> , <i>Alnus glutinosa</i> , <i>A. incana</i> <i>Castanea sativa</i> , <i>Ulmuss elliptica</i> , <i>U. scabra</i> , <i>U. foliaceae</i> , <i>U. suberosa</i>
April	<i>Carpinus orientalis</i> , <i>C. caucasica</i> , <i>Quercus robur</i> , <i>Q. pubescens</i> , <i>Q. petraea</i> , <i>Q. calcarea</i> , <i>Q. iberica</i> , <i>Ulmus laevis</i> , <i>Morus alba</i> , <i>M. nigra</i>
April-May	<i>Pinus hamata</i> , <i>Betula pendula</i> , <i>B. Litwinowii</i> , <i>Fagus</i> <i>orientalis</i> , <i>Quercus macranthera</i>
May-June	<i>Abies</i> , <i>Nordmanniana</i> , <i>Picea orientalis</i>
June-July	<i>Tilia cordata</i> ,

made up on the basis of data supplied by A. A. Grossheim (N 1). The Table shows the different kinds of trees growing in the Trans-Caucasian botanical belt including the investigated region.

As may be seen from this Table, the trees that bloom earliest of all are alders, linden, chestnut, some species of pines and elms. Later, but also during the spring period, bloom the hornbeam, oak, beech, two species of pine trees and elms (*Ulmus* species).

In order to establish that the florescence period of various trees corresponds with the period of their pollen sedimentation on the glacier surface, we carried on a monthly analysis of snow samples taken off the surface of glaciers. The samples were taken in such a way as to include the entire layer of snow accumulated during the month. The data of analysed surface snow samples, for the period from January to May, inclusively, are to be found in Fig. 1, where the composition of the pollen spectrum of trees and concentration of pollen in the sample (per 1 cu. inch of water) are given. The pollen spectra of surface samples reflect the sequence of florescence of various trees indicated in Table N 1: sampling done in April shows a predominance of alder and filbert pollen, in May—of spruce, beech and birch. Fir and spruce flowering takes place in May and June, therefore a maximum amount of pollen of these trees is shed on the glacier surface during the summer period, as is confirmed by the results of analysis of the ice surface sample taken in August. The pollen of grass and bush plants is determined mainly as far as the family possessing on the whole a very long period of blooming, is concerned, for instance: the Gramineae family flowering from May to October, the Chenopodiaceae family—from June to September. The most reliable indication of the season spectrum among this group is presented by the Ericaceae pollen, since the florescence of this systematic group, encountered throughout the West-Caucasian flora belt, takes place in May-July. Among the sporophorous plants the latest period of sporulation—from July to September—belongs to the entire family of ferns Polypodiaceae and Lycopodiaceae. Therefore, their spores will be found in the composition of summer as well as autumn spectra.

In order to draw conclusions on the basis of spore-and-pollen data obtained, it is necessary to remember that the quantity of pollen produced by various plants and their transportability differ considerably. For instance, the polleniferous factor of pine trees is enormous. Furthermore, due to the small size and specific weight of the pollen grain, they can be carried over large distances, therefore, their period of sedimentation is several times larger than the period of florescence, as may be seen in the work of Vareschi (N 4). In almost all our analysed samples the pine pollen predominates or constitutes a considerable percentage, therefore, it cannot be taken as an indication of any particular season. We also took into consideration that fir pollen is produced in great quantities every other year, and will, therefore, show a different percentage in the summer spectra of different years. Owing to its comparatively large specific weight, fir pollen cannot be carried over large distances, similar to that of the pollen, and the period of its sedimentation approaches the period of florescence, which explains its maximal appearance in summer spectra.

Furthermore, in order to make correct conclusions, it is necessary to know also the wind regimen during the entire year. On the southern slope of the Elbrus on the given heights there are prevalent strong south-west and west winds, carrying pollen and spores from remote regions. Therefore surface samples will reveal not only pollen of plants growing in the upper reaches of adjoining valleys, but also pollen brought from the southern slopes of the Main Caucasian Mountain Ridge, such as that of oak trees, spruce, lime, beech, hornbeam. Winds of the following velocity are observed during the year (m/sec.).

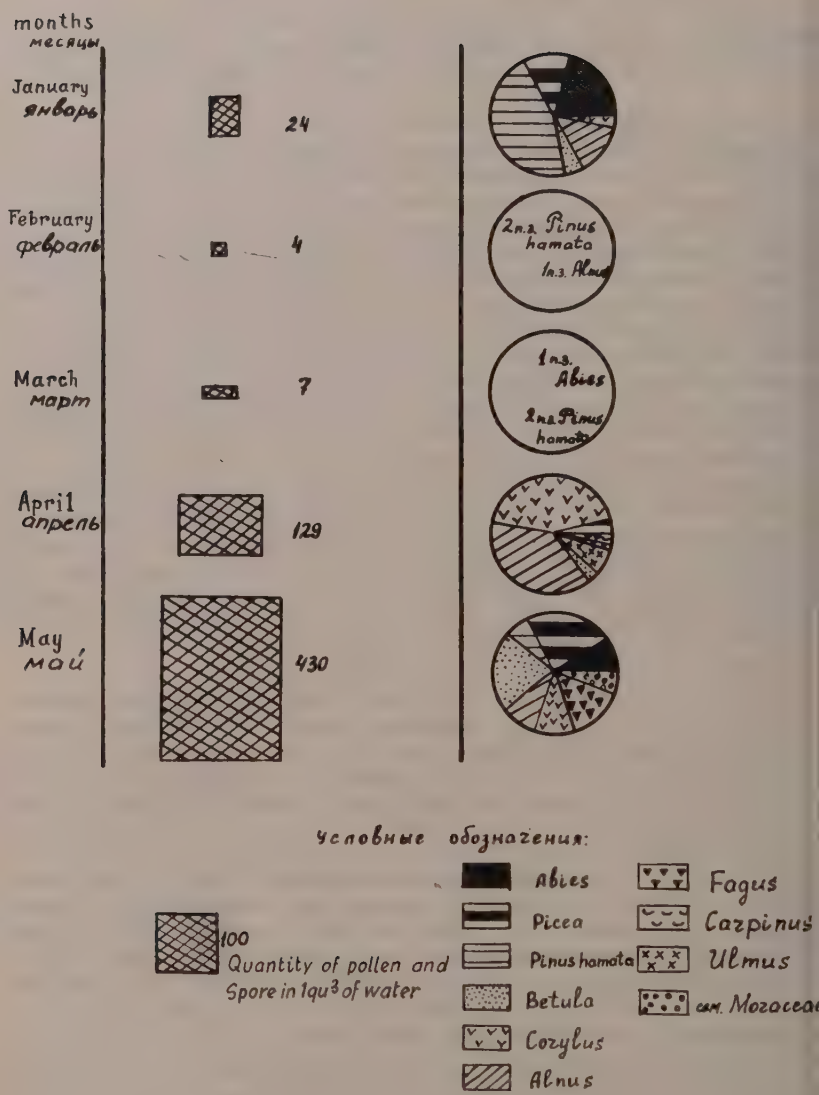


Fig. 1 — The pollen spectrum of the winter and spring surface samples of snow.
Southern slope of altitude 3680 m above sea level.

Winter	Spring	Summer	Autumn	Annual average
7.6	4.9	3.1	6.5	5.5

Minimum wind velocity during spring and summer help to carry and deposit recent sediments of pollen from the air, while in autumn the winds of greater velocity transfer to the glacier larger quantities of pollen, which settled after blooming on the earth surface.

The results of pollen analysis on both glacier crevasses are shown in the diagram giving the composition of the tree pollen, which was adopted as a basis for the season separation (Fig. 2). The spectra of tree pollen are presented in the form of a circle divided into sectors, each of them expressing the percentage content of various species of tree pollen. We believe, that this type of diagram gives the best visual idea of the spectrum as a whole.

Among the pollen spectra of the analysed samples, those for the spring, summer, autumn and winter are shown separately.

The spring sample spectra include NN 3 and 7 of the Garabashi glacier, and NN 10, 8, 5 and 2 from the crevasse near the Refuge of Nine. Predominant in their composition are the pollen of trees blooming in spring, viz: birch, alder, elm, oak, hornbeam, filbert etc. There is a difference between the spring spectra of both glaciers. In the NN 3 and 7 samples from the Garabashi Glacier the predominating pollen is that of alder and birch, constituting about 30-48%, the pollen of broad leaf species: elm, oak, hornbeam, filbert, etc. makes up about 6-8%. Both samples contain a large amount of pollen per 1 cu. inch of water: sample N 3-1214, sample N 7-180. Spring spectra NN 10, 8, 5 and 2 from the Refuge of Nine are distinguished by a higher percent of broad-leaf tree pollen; from 13 to 24% and a comparatively small pollen concentration (up to 50 pollen grain) with the exception of the sample N 10, where up to 22 grains was found.

The samples NN 6 and 12 from the Garabashi glacier and the N 4 from the Refuge of Nine crack may be classed among the spectra of the spring-and-summer transition type. This type of spectrum is distinguished by its approximately similar content of spring and summer season pollen. Thus, the pollen, which is typical for spring spectra makes up a total of 30% in the sample N 12, 13%—in the sample N 6, while the pollen of spruce and fir, mostly produced during the summer season, makes up 22% and 21% respectively in the samples of these tree species. In the sample N 4 the composition predominates mainly in pollen of hornbeam, elm, birch, filbert and alder, blooming in the spring. At the same time, among the herbaceous, there is an obvious predomination of the Ericaceae pollen, while among the spores, the predominant are the fern spores, the maximum of which is, as was already noted above, characteristic for the summer and autumn seasons.

The NN 9 and 4 samples from the Garabashi glacier and NN 1, 7, 9 and 11 from the crack near the Refuge of Nine also belong to the summer spectra. They were separated on the basis of data (Table N 1) concerning the florescence period of various wood species and also as a result of comparing their composition with the data of pollen analysis on summer testing of surface ice, taken in the Ice Basin region during August 1957 (N 3). The composition of this pollen contains over 30% of fir pollen grains. In the pollen spectra of both crevasses classed among the summer seasonal, the highest percentage among the tree species (outside of the pine, which was not taken into consideration for season classing) constitutes the fir pollen grains—up to 54%. The least percentage, though also considerable, was found in the spruce

pollen-from 9 to 13%. Thus, the pollen of these dark-coniferous trees appears predominant in the summer spectra. The latter are not absolutely identical in composition, which is thought to be due to the different quantity of pollen shed by fir trees in different years. For summer samples the high content of tree pollen grains in a volume unit is characteristic, constituting from 146 to 784 pollen grains for the Garabashi Glacier crevasse and from 56 to 174 for the Refuge of Nine crevasse samples.

The samples NN 2, 10 and 11 from the Garabashi crevasse may be counted among the autumn spectra, bearing in mind that stratigraphically they are located directly under the summer spectra, from which they differ in composition. During the autumn period pollen sedimentation takes place on the glacier surface, after having been floating in the air for a considerable time after florescence, as well as sedimentation of pollen, which was lifted off the earth surface by winds. In view of this, it is safe to presume that the composition of autumnal spectra will possess a less definite character, than the summer or spring spectra. The quantity of pollen content in these samples is different: in two of them, in the NN 10 and 11 the concentration is high (600 and 1680 grains in a unit), in the third it is considerably lower—74 pollen grains.

This may possibly be explained by the fact, that during the first half of the fall settlement a larger quantity of pollen takes place than during the second half.

In winter time only solitary grains of pollen settle on the surface of the glacier, which is confirmed by pollen data of winter surface snow samples in our possession (See Fig. 2). The latter serve as criteria for strata separation in the mass of both glaciers. In the Garabashi glacier crevasse we classified as belonging to the winter season the sample spectra: N 1—with 45 pollen grains, N 5-9 pollen grains, while in the Refuge of Nine glacier sample N 6—with 14 grains, N 3-7 grains. As may be seen from the diagram, the winter sample pollen spectra do not possess any qualitative similarity and may be compared only as to quantity.

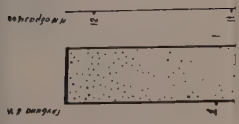
By close sampling in the upper parts of both ice crevasses, there was a consecutive separation of all seasons made, while in the bottom parts of crevasses some seasons were not included in the picture, due to large intervals between adjoining samples.

On the basis of spore-and-pollen analyses it became possible to accomplish a more definite separation of winter, spring and summer strata in both crevasses, in so far as their spectra composition is concerned. Autumn horizons are so far only approximately separated, bearing in mind their stratigraphic position. Pollen spectra of the same seasons taken in both crevasses are similar in composition. The difference between them lies in that the spring and spring-and-summer spectra of the Refuge of Nine crevasse contains a greater amount of broad leaf (hornbeam, elm, beech), than in similar spectra on the Garabashi glacier.

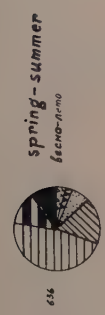
On the whole the spore and pollen content per 1 cu. inch of water in the Garabashi glacier samples is considerably higher than in the glacier near the Refuge of Nine. This is evidently explained by the larger absolute elevation of the latter, and, consequently, by the different meteorological conditions and, above, all, by the wind regimen. As proved by the spore-and-pollen analyses, a large amount of pollen is brought to the glacier surface by south-west and western winds from the Great Caucasian Ridge slopes, including pollen of fir trees, spruce, hornbeams, elms, beech and many other species of trees, which do not grow on the Elbrus slopes.

To explain some of the qualitative and quantitative heterogeneousness of pollen spectra relating to one and the same season and to a better founded segregation of some of them, especially those of the autumn period, it is necessary to carry out additional investigations of the pollen sedimentation conditions during the entire year at different heights.

Thus, it became possible to determine, with the help of the spore-pollen analysis, the time when the snow horizon was formed, and to arrive at the confirmation that



glacier Garabasky



- pine
- birch
- filbert
- alder
- hornbeam
- beech
- elm
- lime
- oak
- snow
- ice
- dirt strip

636 — quantity of pollen grains in 1 qu³ water

crevasse the Refuge of Nine

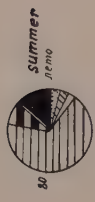
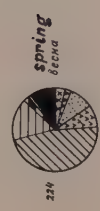
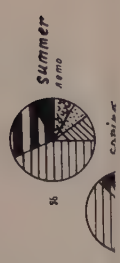


Fig. 2 — The pollen diagrams of the vertical sections of snow-ice layers. The glacier Garabaski and glacier near «Prizut de vjati».

ice bands

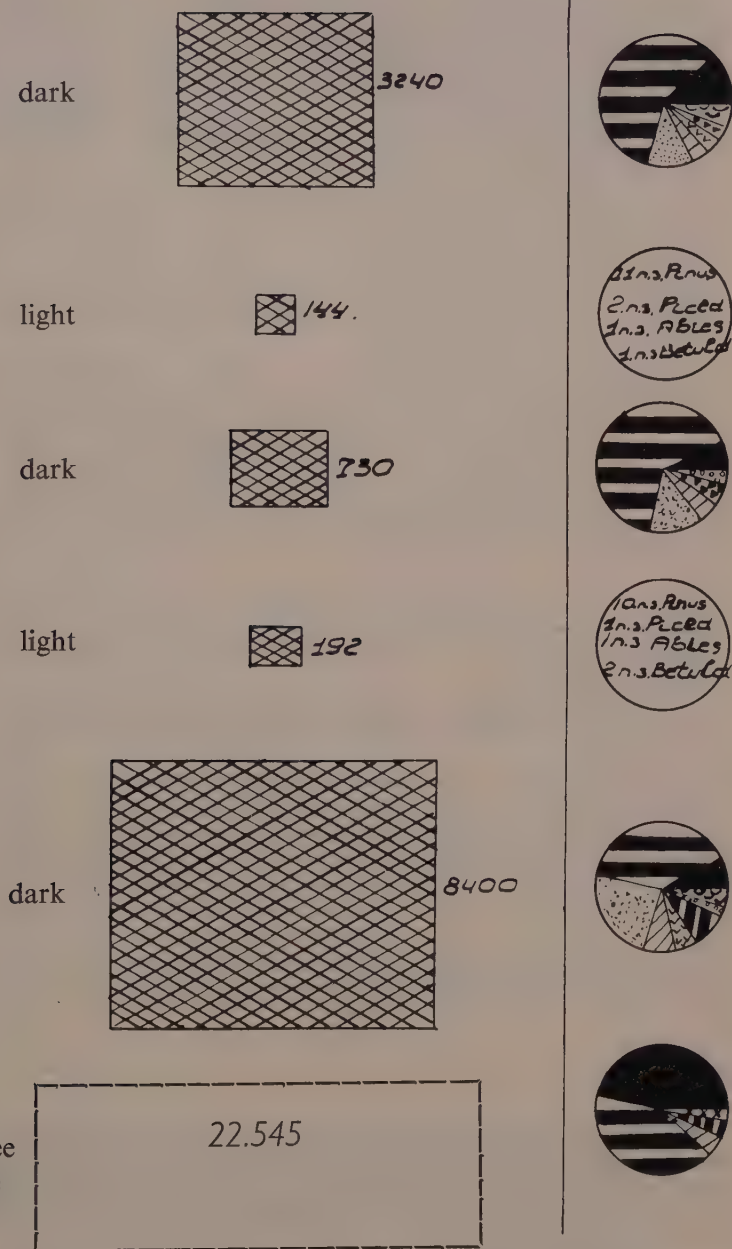


Fig. 3 — The pollen spectrum of dark and light bands of the glacier Kazachaul.

massive polluted ice crusts were formed during summers. On the basis of this it may be accepted, that at a height of 3600-3700 m during the period between summer of 1957 to 1958, the remainder of annual accumulation equalled 190 cm, between the summer of 1956 to 1957—130 cm, between summer of 1955 to 1956—232 cm and between summer of 1954 to 1955—only 100 cm. It may be judged from this, that the accumulation of snow varies considerably from year to year (from 232 cm to 100 cm). Having proved, with the help of the spore-and-pollen analysis, that the stratification of firn masses in the feeding zone is connected with the seasonal accumulation of snow, we then turned to this method to find the origin of striated glaciers.

On the Northern slope of the Elbrus there is the Karachaul glacier descending from the Western peak. The tip of the tongue descends to 3083 m. The glacier surface is undisturbed, crevasses are absent. Plastic flow predominates in it. Lateral, arch-shaped dark bands may be clearly seen on the tongue. The bands are arranged at a distance of 800 m from the tongue. Ice samples for the spore-and-pollen analysis were taken from the dirty and light bands between them. At the same time samples were taken from the surface for summer tests.

The results obtained are shown on Fig. 3, where the concentration and composition of different species of tree pollen is given. The composition of pine pollen, as predominant in these spectra, was excluded, but quantitatively it was included.

An inconsiderable concentration of pollen (144-192) in the sample of light striae was characteristic for these species. In the dark striae samples there were as high as 8400-3240-730 pollen grains. The composition of dark striae pollen spectra including a surface test is quite similar: the pollen of coniferous trees, of spruce and fir trees—predominates, claiming as much as 50%, the surface sample containing over 80%.

Such quantitative and qualitative composition of pollen spectra is characteristic for summer seasons, which is confirmed by the surface test taken in the same region, consequently, there is reason to believe that the dark striae of the glacier were formed in the summer period.

The light striae, containing a minimal quantity of pollen, are evidently, of winter origin and were formed under conditions of minimal sedimentation of pollen on the glacier surface.

Therefore the polluted arch-shaped bands on the glacier surface are nothing but summer horizons formed far back in the firn basin.

However, similar investigations, made on the glacier situated on the southern slope of the Great Asau, did not confirm the dependence between stratification and striation of the glaciers. This is connected with the fact, that the Great Asau Glacier is badly broken up by cracks and ice falls, which causes heavy mixing of the initial substance.

It is, therefore, only in glaciers with plastic flow of arch-shaped dark bands, that the indirect reflection of firn mass stratification may be found.

CONCLUSION

1. With the help of the spore-and-pollen analysis it was made possible to determine the seasons, when separate horizons were formed in the glacier firn basin both in conditions of well-visible stratification, as well as in the absence of same.

2. The size of accumulation on the Southern slope of the Elbrus from 1958 back to 1954 was estimated.

3. The striation of plastic flow glaciers with the stratification of firn masses was established.

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**ETUDES GENERALES SUR LES GLACIERS
GLACIO — METEOROLOGIE**

**GENERAL GLACIER STUDIES
GLACIO — METEOROLOGY**

INVESTIGATIONS GLACIOLOGIQUES, EXECUTEES SUR LE TERRITOIRE D'U.R.S.S. SOUS LE PROGRAMME DE L'ANNEE GEOPHYSIQUE INTERNATIONALE (AGI) en 1957-1959

G. A. AVSIUK

Institut de Géographie, Académie des Sciences, U. R. S. S.

SUMMARY

According to the program of the IGY for 1957-1959, upon the territory of USSR were carried out glaciological investigations in ten areas of glaciation : 1) on the Franz Joseph Land, 2) the Novaya Zemlya, 3) the Polar Urals, 4) the Khibiny Mountains, 5) the Caucasus in the region of Elbrus, 6) the Altai, 7) the Zailiisky Alatau in the Tien-Shan, 9) the Terskei Alatau on the Tien-Shan and 10) the Pamirs on the Fedchenko Glacier. Besides this the glaciation area on the mountain range Kodar, discovered in 1957, was investigated. Besides a great volume of glaciological investigations in the Antarctica was also effectuated. The latter deserves special interpretation and is not touched in this report.

In the majority of the mentioned areas investigations cover two full « Budget glaciological years ». According to the main purpose of the IGY's glaciological works on all the stations was studied a very broad complex of glaciological phenomena, characterizing various forms of the activity of glaciers and their state.

At the most of the stations works on the IGY program was completed towards the autumn 1959. At seven of these stations observations according to a somewhat modified program will continue further.

Primary interpretation of observations, carried out in 1957-1959, will be terminated towards the close of the current year. Nevertheless, it is possible already to state that unique material is gathered, on the base of which it will be possible to forward in an essential way the elaboration of such fundamental problem as the one of the reciprocal action of glaciation and climate; of the evolution of present day glaciation, ancient glaciations and others.

The IGY glaciological investigations brought much new information of zonal and regional features in glaciation phenomena.

Let us cite some examples of formerly unknown features, belonging to glacial phenomena : for instance ablation upon glaciers of the Polar Urals is practically identical upon the entire extension of glaciers and is even somewhat greater in *névé* areas; this is explained by peculiarities of local conditions and the income of warmth.

On the Elbrus wind transfer of snow plays a great role in the feeding of glaciation. This is why the principal snow masses accumulate here in the zone between 4000-4500 m altitude. The *névé* area represents a ring with a break-through in the centre.

On the mountain range Suntar-Khaiata glaciation is represented by two kinds of ice formations — by glaciers as such occupying upper mountain parts and by a ring of constant immovable ice-fields (« *naledi* ») situated at some distance from the glaciers and formed chiefly by newly frozen melt waters from glaciers.

New data were obtained about some original features of autumnal thawing of the Altai glaciers, about the peculiar distribution of a hard and more plastic ice in glacial domes, about the great thickness of soft sediments under glaciers, about the peculiar features of feeding in the New Zemlia glaciers, etc.

These examples illustrate well the variety of glacial phenomena.

The main problems of glaciology may be elaborated with the greatest success only taking into account this variety on the base of total data obtained during the IGY and in the united work with this aim of glaciologists belonging to different countries.

RÉSUMÉ

Selon le programme de l'A.G.I. pour 1957-1959, l'on a exécuté sur le territoire d'U.R.S.S. des investigations glaciologiques dans 10 régions de glaciation : (1) sur la terre de François-Joseph; (2) sur la Novaya Ziemla; (3) sur les monts de l'Oural Polaire; (4) dans les Khibiny; (5) au Caucase dans la région de l'Elbrouz; (6) sur l'Altai; (7) sur la chaîne de montagne Suntar-Khayata; (8) sur l'Alataou Transilien du Thian-Chan; (9) sur le Terskoy-Alataou du Thian-Chan et (10) dans le Pamir sur le glacier Fedtchenko. En outre, fut aussi investiguée la région de glaciation dans

la chaîne de montagne Kodar, découverte en 1957. En dehors de cela, l'on a aussi exécuté un grand volume de travaux glaciologiques dans l'Antarctide. Ces derniers méritent une interprétation spéciale et ne sont point approchés dans l'exposé donné.

Dans la majorité des régions mentionnées, les investigations embrassent deux « années glaciologiques de budget » complètes. En accord au but principal des travaux glaciologiques sous l'A.G.I., l'on a étudié sur toutes les stations un très vaste complexe de phénomènes glaciologiques, qui caractérise des formes variées de l'activité vitale des glaciers et leur état.

Sur la majorité des stations, les travaux sous le programme de l'A.G.I. furent achevés en automne 1959. Sur sept d'entre elles les observations vont continuer sous un programme quelque peu modifié.

L'on exécute le traitement primaire des observations de 1957-1959; il sera achevé vers la fin de l'année courante. Néanmoins l'on peut déjà soutenir que le matériel accumulé est unique, à base duquel il sera possible de s'avancer essentiellement dans l'élaboration entre la glaciation et le climat, l'évolution de la glaciation actuelle, les glaciations anciennes, etc.

Les investigations glaciologiques, sous l'A.G.I., ont donné beaucoup d'informations nouvelles sur les particularités zonales et régionales des phénomènes glaciaires.

Citons quelques exemples de particularités des phénomènes glaciaires, inconnus auparavant, par exemple, l'ablation sur les glaciers des monts de l'Oural Polaire est, en pratique, presque pareille sur toute l'étendue des glaciers et elle est même plus grande dans les régions du névé; ceci s'explique par les particularités de l'alimentation et de l'apport d'air chaud.

Sur l'Elbrouz le transport éolien de la neige joue un grand rôle dans l'alimentation de la glaciation. Voici pourquoi ici les masses principales de la neige s'accumulent dans une zone entre des altitudes de 4000 à 4500 m. La région de névé représente une manière de coral, déchiré au milieu.

Dans la chaîne de montagne Suntar-Khayata, la glaciation est représentée par deux aspects de formations glaciaires : par des glaciers comme tels, qui occupent les parties supérieures des montagnes, et par un cercle de « nalédys » constantes, qui se trouvent à une certaine distance des glaciers et qui sont formés, au fond, par des eaux gelées de fonte des glaciers.

L'on a obtenu des nouvelles données sur les particularités de la fonte d'automne des glaciers de l'Altai, sur la distribution originale de la glace dure et quelque peu plus plastique dans les couloirs glaciaires, sur la grande puissance des dépôts meubles sous les glaciers, sur les particularités de l'alimentation des glaciers de la Novaya Zemla, etc.

Ces exemples illustrent bien la diversité des phénomènes glaciaires.

Les grands problèmes de glaciologie peuvent être élaborés avec le plus de succès seulement en tenant compte de cette diversité, à base de tout le total des données obtenues de l'A.G.I. et sous l'union en ce but des glaciologues de maints pays. L'on se représente que ceci est la voie la plus correcte, qui assure un développement nécessaire de la glaciologie.

La majorité de rapports concernant les études des glaciers représentés par des glaciologues soviétiques à la XII^e Assemblée de l'Union Géodésique et Géophysique Internationale sont fondés par excellence sur des données obtenues dans la période de l'Année Géophysique.

A cet effet on trouve rationnel de faire un aperçu général des travaux de recherches glaciologiques exécutés par l'Union Soviétique selon le programme de l'AGI. Ces travaux ont été menés depuis 1957 sans interruption jusqu'à 1959 et, ayant le caractère complexe, ont largement embrassé de nombreux aspects des phénomènes glaciaires.

Les travaux glaciologiques ont été faits systématiquement de 1957 à 1959 dans dix régions glaciaires dans de différentes parties de l'Union Soviétique. Les glaciologues soviétiques ont également réalisé de grands travaux glaciologiques en Antarctide. Ces derniers méritent une communication à part et c'est pourquoi l'on n'en parlera point dans le présent exposé.

Les travaux de recherches systématiques ont été réalisés en Union Soviétique sur la base des stations glaciologiques installées dans les régions suivantes :

- 1) l'Archipel de François Joseph, la baie Tikhaia de l'île Gouker;
- 2) l'île au Nord de Novaia Zemlia, le port Rousski;
- 3) Oural Polaire;

- 4) Khibiny;
- 5) Caucase, la région d'Elbrous;
- 6) Altai;
- 7) la chaîne Sountar-Khaiata (Yakoutie orientale, l'amont d'Indigirka);
- 8) Thian-Chan, la chaîne Zailisky Alataou;
- 9) Thian-Chan, la chaîne Tersky Alataou;
- 10) Pamir, le glacier de Fedtchenko.

Les coordonnées des stations mentionnées sont publiées dans la liste des stations de l'AGI.

Les études principales sur l'archipel de François Joseph étaient effectuées sur la nappe glaciaire de l'île Gouker, notamment sur les coupoles glaciaires de Tchourlianis et de Djekson, ainsi que sur les glaciers de Sédov et d'Elena. Pour réaliser des observations systématiques sur le sommet de la coupole Tchourlianis ont été construites deux maisons, où ont vécu et travaillé des glaciologues pendant 1957-1959. Une pareille maison a été construite sur le glacier de Sédov, ainsi que de petites cabanes sur la coupole de Djekson et sur le glacier d'Elena. La station centrale a été installée dans la baie Tikhaia. En dehors de ces travaux systématiques réalisés dans des endroits caractéristiques sur quelques glaciers de l'île Gouker on faisait des observations périodiques et a réalisé des «travaux d'itinéraire».

Outre les travaux sur l'île de Gouker on a effectué quelques observations sur des glaciers des îles Heisse et Scotte-Kelti et a examiné de l'avion les glaciers de la partie centrale de l'archipel.

Les travaux sur l'archipel de François Joseph ont été faits sous la direction de V.L. Soukhodrovski.

A la Novaia Zemlia des travaux principaux se sont effectués sur la nappe glaciaire de l'île du Nord dans la région du glacier de Chokalski descendant dans la baie d'Otkoupchtchikov, dans la région du Port Rousski et sur le glacier proprement dit. En vue des travaux systématiques qui devaient avoir lieu sur toute l'étendue entre la côte et la ligne de partage des eaux (entre la mer Barentsov et la mer de Karsk) on a construit trois maisons : l'une sur la nappe glaciaire à la distance de 40 kilomètres environ du bord, l'autre sur le dit barrage de Doutes — là, où pendant la deuxième API (1932-1933) ont été effectuées les travaux d'observation, et, enfin, la troisième — sur la partie inférieure du glacier de Chokalsky. La base principale a été installée au bord de la mer. Ici, comme sur l'archipel de François Joseph, les travaux essentiels ont été complétés par les travaux périodiques et les «travaux d'itinéraire».

Les travaux sur la Novaia Zemlia ont été dirigés par N. M. Svatkov et O. P. Tshijov.

Dans la région de l'Oural Polaire on travaillait sur les glaciers de l'Institut de Géographie, d'Obroutchev et de l'Université de Moscou. Sur les deux premiers ont été bâties des maisons spéciales. La base centrale est située près du lac Khodatac. Les directeurs des travaux sont R. I. Venieri et L. S. Troitski.

Les travaux glaciologiques dans les régions de l'archipel François Joseph, de Novaia Zemlia et d'Oural Polaire ont été réalisés par des collaborateurs de l'Institut de Géographie de l'Académie des Sciences de l'URSS sous la direction générale du Professeur G. A. Avsiuk.

La plupart des travaux dans Khibiny a été consacrée à la nappe de neige et aux avalanches dans la région du plateau Ioukspor et sur un petit glacier naissant seul, connu dans cette région qui doit présenter pour les glaciologues un intérêt particulier.

Au Caucase on travaillait surtout sur les glaciers d'Elbrous. Ce travail systématique a été possible grâce à ce qu'on a disposé des stations spéciales précédemment construites et des installations alpinistes. En dehors des travaux systématiques menés sur Elbrous même on a effectué des travaux de recherches sur quelques autres glaciers du Caucase.

Les travaux à Khibiny et sur Elbrous ont été réalisés par des collaborateurs de la Faculté géographique de l'Université de Moscou sous la direction du Professeur G. K. Touchinski.

Dans la région d'Altai le travail systématique a été fait dans la partie sud-est de l'Altai Central sur les glaciers Grand et Petit Aktourou et sur le glacier Taldourinski. On a également étudié plusieurs autres glaciers d'Altai. Le travail a été réalisé par des collaborateurs de l'Université de Tomsk sous la direction du Professeur M. V. Tronov.

Dans la chaîne Sountar-Khaiata (Yakoutia orientale) qui représente un contrefort des montagnes Verkhoianskie des glaciers proprement dits, des «naledis», des sols congelés, le matériel rocheux et certains phénomènes géocriologiques constituaient des objets principaux des recherches. On examinait systématiquement le glacier n° 31, certains «naledis» et les terrains libres de la glace. Là-dessus on a réalisé un grand nombre de recherches «d'itinéraire». Sountar-Khaiata représente un intérêt particulier comme la région de haute montagne située à proximité du pôle du froid de l'hémisphère boréal de la Terre.

Les travaux sur Sountar-Khaiata ont été exécutés par des collaborateurs de l'Institut de Congélation de l'Académie des Sciences de l'URSS sous la direction de N. A. Grave.

Dans la région de Thian-Chan (chaîne de Zailiski Alataou) on a réalisé des recherches systématiques sur les glaciers de Touiouksou dans le bassin du fleuve Malaia Almatinka. Pour la réalisation des travaux de recherche systématiques sur toute l'étendue du glacier de Touiouksou on a construit dans 3 zones : dans celle de l'amont, sur la langue et sur la moraine terminale, trois maisons pour le travail et le logement. En dehors de cela on a mené des travaux périodiques sur d'autres glaciers dans le bassin de Malaia Almatinka. On a également réalisé des recherches «d'itinéraire» importantes sur toute une série de glaciers. Le travail a été effectué par des collaborateurs de la Section de Géographie de l'Académie des Sciences de la RSS de Kazakstan sous la direction de K. G. Makarevitch. A certains travaux ont pris part des glaciologues de la République Démocratique Allemande et des collaborateurs de l'Institut de Géographie de l'Académie des Sciences de l'URSS.

On a travaillé également à une autre station installée dans la région de la chaîne Terski Alataou (bassin du fleuve Tchou-Kzyl-Sou).

L'objet principal des études systématiques a été le glacier Karabatkak. Sur ce glacier on a étudié systématiquement depuis 1948 certains phénomènes glaciaires. On a également réalisé les études glaciologiques périodiques et «d'itinéraire» sur plusieurs glaciers de la chaîne Terski Alataou de la zone de «syrte» (dans les chaînes Kok-Chaal, Borkoldoi, Djetymbel, Kouiliou, Sarydjas, Akchiriak et autres). Au cours des travaux «d'itinéraire» on a fait plusieurs levés phototéodolites. La base principale de tous ces travaux se trouvait à la station de haute montagne de Thian-Chan construite en 1947.

Les travaux ont été réalisés par des collaborateurs de la Section de Géographie de l'Académie des Sciences de la RSS de Kirgizie sous la direction de R. D. Zabirow, avec la participation des collaborateurs de la Faculté géographique de l'Université de Moscou et de l'Institut de Géographie de l'Académie des Sciences de l'URSS.

Dans la région de Pamir on travaillait sur le glacier de Fedtchenko, l'un des plus grands glaciers montagneux de l'Union Soviétique et du monde entier. Les recherches systématiques ont été effectuées à deux stations spécialement construites : la première à proximité de l'aval du glacier à l'altitude de 3000 m environ (chef de la station L. P. Tribounski), la deuxième sur le névé du glacier à l'altitude de 5000 m (chef de la station V. K. Nozdrioukhine). Les conditions de vie et de travail ont été particulièrement difficiles et compliquées à la deuxième station, qui est peut-être l'une des plus hautes stations ayant fonctionné sans interruption durant toute l'Année Géophysique.

sique Internationale. Cette petite maison d'aluminium, installée directement sur le névé et emportée par le glacier en mouvement, a fait au cours de l'Année Géophysique Internationale un trajet de plusieurs centaines de mètres. A maintes fois sous la maison la glace se brisait en formant des crevasses; en hiver elle a été enterrée sous la neige. En dehors de ces deux stations dans la période de l'Année Géophysique Internationale sur le glacier de Fedtchenko a continué à fonctionner la troisième station installée dans la partie moyenne du glacier à l'altitude de 4000 m, construite déjà à l'époque de la deuxième Année Polaire Internationale. Dans la période de l'Année Géophysique Internationale on a également réalisé des travaux périodiques dans différentes parties du glacier de Fedtchenko et sur quelques glaciers-affluents. On y a aussi fait des travaux «d'itinéraire». Les travaux ont été effectués par l'Expédition glaciologique de l'Institut de Mathématique et de Mécanique de l'Académie des Sciences de la RSS d'Ouzbekistan sous la direction du Docteur en sciences physiques et mathématiques V.I. Goubine, M.A. Petrosiants et V.F. Souslov. Au travail ont pris part des collaborateurs de l'Institut de Physique de la Terre de l'Académie des Sciences de l'URSS, des Universités de Moscou et de Léninegrad, ainsi que des glaciologues de la République Démocratique Allemande, de la République populaire Chinoise et de la Pologne.

Afin d'obtenir des données comparatives dans la période de l'Année Géophysique Internationale ont été effectuées des études systématiques de la couverture de neige, du bilan thermique et du chargé thermique dans la couche supérieure de lithosphère en dehors des régions des glaciers contemporains — sur la station Zagorsk aux environs de Moscou où on s'occupait en même temps des problèmes de méthode et réalisait des travaux expérimentaux. La majorité de ces études a été faite par des collaborateurs de l'Institut de Congélation de l'Académie des Sciences de l'URSS; des collaborateurs de l'Institut de Géographie de l'Académie des Sciences de l'URSS y ont également participé. La direction générale a été réalisée par les docteurs en sciences géographiques I.Y. Baranov et A.V. Goloubev.

Parallèlement aux recherches scientifiques accomplies aux stations mentionnées ci-dessus toute une série de travaux glaciologiques a été réalisée dans la période de 1957-1959 dans diverses régions de congélation; par exemple : dans la région de congélation dans la chaîne Kodar (située à l'Est du lac Baikal) découverte en 1957 et auparavant inconnue; sur les glaciers de la chaîne Djoungarski Alataou, dans plusieurs régions glaciaires de Caucase (excepté celle d'Elbrous) et dans beaucoup d'autres endroits de notre pays.

Les études accomplies à la majorité des stations de l'Année Géophysique Internationale ont entièrement embrassé deux «années glaciologiques budgétaires» (1957-1958 et 1958-1959) aux caractères très différents, ce qui représente un intérêt particulier. Les travaux prévus dans le programme de l'AGI ont été achevés en automne de 1959 (après la période d'ablation) à la plupart des stations, le reste a été terminé le 31 décembre 1959.

Actuellement 3 stations (à l'archipel de François Joseph, à la Novaia Zemlia et sur Sountar-Khaiata) ne fonctionnent plus; les autres continuent leur travail suivant un programme un peu changé, souvent sensiblement réduit.

Dans la période de l'AGI la coordination générale des travaux a été réalisée par le Groupe de Travail Glaciologique du Comité Soviétique de l'AGI (Secrétaire Général du Groupe — Prof. G.A. Avsiuk). A l'avenir la coordination du travail sera probablement concentrée dans la Section de Glaciologie du Comité Soviétique de Géodésie et de Géophysique.

Pour la meilleure unification des travaux au cours de l'Année Géophysique Internationale le Groupe de Travail Glaciologique a élaboré «Programme et instructions générales pour la réalisation du travail glaciologique de l'Année Géophysique Internationale», ainsi que «Instructions principales pour différents aspects du

travail glaciologique (15 éditions) qui ont été publiés par le Comité Soviétique de l'Année Géophysique Internationale en 1957 et envoyés à toutes les personnes s'occupant des recherches glaciologiques. Grâce à cela on a réussi d'obtenir une certaine uniformité des programmes et des méthodes de travail bien qu'encore insuffisante. En tout cas c'est déjà un succès.

Le but principal des travaux glaciologiques de l'Année Géophysique Internationale est l'étude des phénomènes d'accumulation de transformation, de mouvement et fonte de la glace sur la surface du sol dépendamment du bilan thermique de la surface du sol, ainsi que l'étude du rôle des glaciers dans la circulation de l'eau sur notre planète.

Les problèmes les plus importants qui doivent être envisagés sur la base des données de l'Année Géophysique Internationale sont les suivants : l'action réciproque des glaciers et du climat; l'état actuel, la répartition, la puissance de la congélation et le sens de son évolution actuelle; les particularités zonales et régionales du développement de la congélation et la manifestation des phénomènes glaciaires. A tous ces problèmes est étroitement lié celui des glaciers anciens et surtout leur dynamique et leur influence sur l'évolution de la nature de la Terre.

Afin d'obtenir des matériaux nécessaires pour l'étude de ces problèmes il a été indispensable de réaliser des études complexes concernant un très grand nombre de questions.

A cet effet les travaux de l'Année Géophysique Internationale visant à obtenir des données relatives à l'état spatial, à la puissance et à la tendance de développement des glaciers prévoient en même temps l'étude de l'alimentation et la fusion dans les glaciers, la transformation des matériaux accumulés en névé et en glace, les métamorphoses de la glace, la migration de l'eau dans la glace et dans la neige, le mouvement de la glace dans les glaciers, le régime thermique et le bilan thermique des glaciers, l'échange thermique entre la surface du glacier et l'atmosphère, les propriétés physiques et mécaniques de la glace et de la neige, la structure des glaciers, l'âge des couches de névé et de glace, l'écoulement des eaux de fonte, la formation des dépôts glaciaires, le travail d'érosion produit par la glace, les phénomènes géocriologiques etc.

L'étude de tous ces phénomènes a été possible grâce à l'organisation des travaux conformes météorologiques, hydrologiques, glaciologiques, géophysiques, géomorphologiques et autres, au collectionnement des échantillons, ainsi qu'à toute sorte de travaux analytiques.

Il faut remarquer que dans la période de l'AGI on appliquait beaucoup plus largement qu'auparavant des méthodes précises telles que : méthodes électrométriques, séismométriques, photogrammétriques, le forage de la glace, etc.

Au moment actuel on commence à travailler l'immense quantité de données obtenues pendant 1957-1959; ce travail ne sera achevé que vers la fin de l'année courante, c'est pourquoi il est encore tôt de parler de conclusions scientifiques.

Néanmoins il est possible d'affirmer à partir d'ici que la documentation recueillie garantit des succès fondamentaux dans le domaine de la glaciologie, l'ampleur des études glaciologiques, leur durée et leur simultanéité étant sans précédent. On peut sans exagérer dire que l'Année Géophysique a créé une nouvelle étape dans le développement de la science glaciologique.

Il est évident que sur la base des données glaciologiques de l'Année Géophysique Internationale et des matériaux des études météorologiques, océanologiques et certains autres on pourra envisager la possibilité de la réalisation des projets grandioses comme par exemple, l'élimination de la couverture glaciaire de l'Océan glacial du Nord.

Des études glaciologiques de l'Année Géophysique ont donné beaucoup de nouveaux matériaux concernant des particularités régionales et zonales des phénomènes glaciaires ce qui est très important non seulement pour comprendre des règles

locales de la congélation (et en profiter conformément), mais aussi pour aboutir à la synthèse sans faute.

On peut dès maintenant, sans attendre qu'on finisse la classification des matériaux obtenus, citer quelques exemples des phénomènes glaciaires particuliers ayant lieu dans différentes régions glaciaires.

L'ablation des glaciers de l'Oural Polaire est pratiquement la même sur toute leur étendue et même elle est plus sensible dans les régions de névé que sur les langues. Ce «paradoxe» résulte des particularités de leur alimentation «par le vent», du caractère et de la répartition de la chaleur — ces glaciers fondent sous l'action plutôt de la chaleur advective que de la chaleur radiale et, partiellement, sous l'action de la chaleur des foehns et des dépôts liquides. Par suite du fait que l'alimentation «par le vent» représente pour ces glaciers l'alimentation essentielle, le régime de température de la masse glaciaire ne correspond pas au climat rude continental de cette région et s'approche du type contraire du régime de température — le régime des glaciers dits «tièdes» propres au climat de mer.

Dans la région d'Elbrous (Caucase) qui par sa nature diffère sensiblement de l'Oural Polaire, le transport de la neige par le vent joue également un rôle important dans l'alimentation de congélation. Dans le plan la configuration de cette dernière ressemble même à une certaine mesure à la «rose des vents». Grâce au transport par le vent les masses principales de neige s'accumulent dans l'intervalle entre 4000 et 4500 mètres. Dans des régions plus hautes par endroits il n'existe pas de névé. Là, la zone de névé constitue une sorte de cercle interrompu dans la partie centrale. Les notions de la limite climatique et de la limite des neiges de saison n'existent pas pour Elbrous; quant à la répartition des régions d'alimentation et de fonte elle est très originale et variée.

Suivant les données de la deuxième Année Polaire Internationale de 1932-1933 le bouclier glaciaire de l'île septentrionale de Novaia Zemlia est privé de couverture de névé. A cause de cela on pensait que la congélation de Novaia Zemlia n'a pas actuellement de source d'alimentation et représente une formation reliquat disparaissante. Les études faites au cours de l'Année Géophysique Internationale, ainsi que certains travaux précédents ont démontré l'erreur de pareilles conclusions. Il est possible qu'à l'époque de la Deuxième Année Polaire Internationale l'épaisseur de la névé a été plus petite et localement n'a point existé. S'il en est ainsi, il est possible qu'au cours des 25 années passées depuis la Deuxième Année Polaire Internationale l'alimentation du glacier de Novaia Zemlia s'est améliorée. A ce problème est consacrée la communication de N. M. Svatkov, présentée à la XII^e Assemblée de l'UGGI.

Dans la région de l'archipel de François Joseph la différence relativement peu sensible entre les altitudes des glaciers suivie de la différence climatique également petite provoque la formation des masses glaciaires aux régimes de température et aux phénomènes glaciaires assez différents. Ces problèmes représentant un grand intérêt sont envisagés dans l'exposé de A. N. Krenke.

Dans la chaîne Sountar-Khaiata la congélation contemporaine s'exprime dans deux types fondamentaux de formations glaciaires : les glaciers proprement dits occupant les sommets des montagnes et le cercle de grands «naledis» permanents (plusieurs kilomètres de longueur) éloignés des glaciers à une certaine distance et disposés sur le niveau hypsométrique plus bas.

Il faut remarquer que les glaciers représentent des formations sédimentaires dues aux précipitations atmosphériques solides, tandis que les «naledis» sont des corps glaciaires hydrogènes, constituées généralement par des eaux de fonte regelées. La combinaison de ces deux types de formations glaciaires résulte du climat foncièrement continental et caractérise la congélation actuelle de cette région.

Avant l'AGI on évaluait l'épaisseur des masses glaciaires dues à la congélation

moderne ainsi qu'à l'ancienne habituellement selon les symptômes indirects et on l'estimait beaucoup moins importante qu'elle ne l'était.

Par exemple, on a obtenu selon des données des études sismiques réalisées au cours d'AGI (confirmées parfois au moyen de forage et de mesures électrométriques) des valeurs suivantes de l'épaisseur de glace : le glacier de Fedtchenko (Pamir), la partie moyenne — 800 m environ, la partie basse — 250 m environ, chiffre maximum — 900 m; le glacier de Toubioksou (Thian-Chan, Zailiski Alataou) — l'amont — 100-140 m, la partie moyenne — 60-80 m, la région d'aval — 30-40 m. La puissance de la glace de certains glaciers d'Altai atteint 340 m, tandis que l'épaisseur moyenne des glaciers d'Altai dépasse évidemment 100 m.

Les études sismiques de l'épaisseur des glaciers de Thian-Chan, de Pamir et d'Altai ont donné des résultats un peu inattendus pour l'épaisseur du matériel friable (à ce qu'il paraît morainique), renfermé entre la surface inférieure de la glace et le lit rocheux du glacier. Ainsi dans les glaciers de Zailiski Alataou (Thian-Chan) la puissance du matériel friable couché sous la glace se limite par 160 m; dans le glacier de Fedtchenko (Pamir) elle atteint 400 m et dans les glaciers d'Altai, elle dépasse évidemment 100 m. L'origine de ces dépôts friables n'est pas encore connue, néanmoins il est évident qu'ils ne proviennent pas de l'activité des glaciers modernes. La présence sous la glace en mouvement du matériel friable fait reviser la notion du travail d'exaration et de transport des glaciers montagneux, ainsi que certaines questions relatives à l'histoire de la congélation de ces pays.

Ces quelques exemples illustrent bien la variété des phénomènes glaciaires propres aux régions continentales de la congélation contemporaine. L'étude de la diversité et des particularités, des phénomènes glaciaires, ainsi que l'analyse des traits spécifiques à leur combinaison constituent, à notre avis, le but essentiel et le plus intéressant. Nous estimons que c'est à la condition seule de l'examen méticuleux des particularités zonales et régionales de l'évolution de congélation qu'on puisse avancer dans la compréhension des problèmes principaux de l'évolution de la congélation de la Terre.

On peut citer comme l'une des conclusions générales déjà suffisamment claire la constatation de la réduction de la congélation dans toutes les régions où on a effectué des travaux glaciologiques durant 1957-1959. Néanmoins on a constaté l'extension de quelques glaciers provoquée par des causes locales éventuelles propre à la congélation dans la période de la régression générale. Dans une série de régions glaciaires (telles que : Caucase, Novaia Zemlia et certaines autres) on a constaté l'accentuation de l'alimentation au cours des dernières années. Cela fait supposer qu'il est possible que la retraite contemporaine des glaciers doit s'arrêter à l'avenir prochain et même sera remplacée par la période de leur développement. Cependant nous croyons qu'il est prématuré de faire de pareilles conclusions avant que le travail sur des données d'AGI ne soit accompli.

Au moment donné une série d'ouvrages a déjà apparu basés sur les matériaux d'AGI. La majorité d'eux est publiée dans des publications consacrées à l'AGI (bulletins et recueils de travaux de l'AGI, «Matériaux» et «Résultats» des études glaciologiques d'AGI, revues scientifiques). Un nombre déjà considérable de questions glaciologiques bien qu'encore particulières est traité sur la base des matériaux d'AGI dans les communications présentées à la XII^e Assemblée de l'UGGI.

Le développement des recherches glaciologiques doit consister : premièrement dans l'assimilation thématique des résultats de l'AGI — l'étude des problèmes régionaux, ainsi que la mise au point des problèmes fondamentaux de la glaciologie; deuxièmement dans la réalisation future des études glaciologiques sur le terrain au niveau scientifique qui soit aussi haut que celui de l'AGI. Il est également indispensable de continuer le travail du perfectionnement des méthodes glaciologiques.

Il va sans dire que la mise au point des problèmes fondamentaux de la glaciologie ne peut être accomplie que sur la base de l'ensemble de matériaux de l'AGI.

et à condition de la participation des glaciologues de divers pays, c'est-à-dire, sur la base de la collaboration scientifique internationale.

Nous croyons rationnel que l'union glaciologique internationale concentre ses efforts sur l'organisation prochaine des études scientifiques internationales du problème — «Dimensions, régime et dynamique de la congélation moderne comme résultats de son action mutuelle avec le relief et le climat». Une telle orientation de l'activité de l'union glaciologique internationale donnerait, estime-t-on, un grand effet scientifique indispensable à résoudre toute une série de questions de la pratique actuelle.

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GLAZIOLOGISCHER KOMMENTAR ZUR NEUEN IM HERBST 1957 AUFGENOMMENEN KARTE 1: 10.000 DES GROSSEN ALETSGLETSCHERS

P. KASSER (Switzerland)

Abteilung für Hydrologie und Glaziologie der Versuchsanstalt für Wasserbau und
Erdbau an der Eidg. Technischen Hochschule, Zürich

SUMMARY

The accompanying map «Aletschgletscher, Blatt 3, 1:10'000, Stand 1957» is the first of 4 sheets that together embrace the entire drainage basin of the Great Aletsch Glacier. The following commentary describes briefly the purposes of the mapping, the methods of photogrammetry and cartography, and the possibilities for application of the finished map. Section 4, figures 3 and 4, and table 1 contain some preliminary values of the changes in elevation of the glacier surface between the years 1851, 1926/27, 1947 and 1957, and also of the volume decrease of the glacier between 1926/27 and 1957.

RÉSUMÉ

La carte annexée «Aletschgletscher, Blatt 3, 1:10'000, Stand 1957» est la première des 4 feuilles qui comprennent le bassin versant complet du Grand Glacier d'Aletsch. Ce commentaire traite brièvement le but, le relevé, l'élaboration et les possibilités d'application de cette carte. Chapitre 4, figures 3 et 4 et tableau 1 contiennent quelques chiffres préliminaires sur la variation en altitude de la surface du glacier entre 1851, 1926-27, 1947 et 1957 ainsi que sur la diminution en volume depuis 1926/27.

ZUSAMMENFASSUNG

Die beiliegende Karte «Aletschgletscher, Blatt 3, 1:10'000, Stand September 1957» ist das erste der 4 Blätter, welche das gesamte Einzugsgebiet des Grossen Aletschgletschers umfassen. In diesem Kommentar wird über Zielsetzung, Aufnahme, Ausführung und Anwendungsmöglichkeiten dieses Kartenwerkes kurz berichtet. Einige vorläufige Zahlen für die Höhenänderung der Gletscheroberfläche zwischen 1851, 1926/27, 1947 und 1957, sowie über die Volumänderung zwischen 1926/27 und 1957 sind in Kapitel 4, in den Bildern 3 und 4 in Tabelle 1 enthalten.

1. EINLEITUNG

Erstmals erscheint der Name Aletsch auf der Karte, die Antoni Lambien [5] (*) im Jahre 1682 herausgegeben hat. Doch erst mit den Kartenwerken des 19. Jahrhunderts wurden brauchbare Grundlagen für glaziologische Studien geschaffen. Zwischen 1832 und 1864 erschien die Dufourkarte (**) der Schweiz, in Schraffenmanier und im Masstab 1:100'000 ausgeführt, zwischen 1880 und 1900 der Siegfriedatlas (***), eine Kurvenkarte, die im Gebirge einen Masstab von 1:50'000 und 30 m Aequidistanz aufweist. Für diese beiden Kartenwerke wurden die Vermessungen im Aletschgebiet um 1850 herum, zur Zeit des letzten Gletscherhochstandes, durchgeführt. J. R. Stengel arbeitete 1851 in der Gegend des Konkordiaplatzes. Die erste photogrammetrische Aufnahme, ausgewertet im Masstab 1:25'000 mit 20 m Kurvenabstand und publiziert im Masstab 1:50'000 als Landeskarte Blatt 264, erfolgte terrestrisch in den Jahren 1926/27. Ein Teil des Gebietes, nämlich Ewigschneefeld, Jungfraufirn und die Zunge des Grossen Aletschgletschers wurden 1947 luftphotogrammetrisch aufgenommen und im Masstab 1:25'000 ausgewertet.

(*)vgl. Literaturnachweis am Ende der Arbeit

(**) Guillaume—Henri Dufour 1787-1875.

(***) Hermann Siegfried 1819-1879.

Ungefähr gleichzeitig mit dem Entstehen brauchbarer Karten setzten am Aletsch-
gletscher auch die glaziologischen Studien ein (Agassiz und Desor 1841, Tyndall
noch vor 1850, Grad und Dupré 1869). Im Rahmen der durch die Schweiz. Gletscher-
commission (F. A. Forel) im Jahre 1880 begonnenen Beobachtungen an den Gletscher-
enden wird die Lageänderung des Gletschertores seit 1892 beobachtet.

1908-1914 untersuchte Lutschg das Verhalten des Märjelsees. 1921 kam die
neute noch dienende Abflussmesstation mit Limnigraph in Betrieb. 1929 bestimmten
Mothes und Sorge mittels Seismik eine maximale Eismächtigkeit von rund 800 m
im Konkordiaplatz. Zwischen 1930 und 1940 führten Vareschi pollenanalytische
Studien im Ablationsgebiet und Seligman Firnstudien am Jungfraufirn durch. 1948
bohrte Perutz im Jungfraufirn, um die Geschwindigkeitsverteilung längs einer Verti-
kalen zu messen.

Ab September 1918 wird der Firnzuwachs auf dem Jungfraufirn kontinuierlich
beobachtet; seit 1941 werden die periodischen Messungen auf breiter Basis ausgebaut.
Das Netz umfasste ab 1942 den ganzen Jungfraufirn, schloss ab 1945 den Konkordi-
aplatz ein und erstreckte sich ab 1949 bis Märjelen, ab 1950 bis zu den Katzlöchern
und seit 1951 auf den gesamten Hauptstrom vom Jungfraujoch bis zum Gletschertor
4,6,7,8].

Bei diesen Arbeiten hat sich immer mehr das Bedürfnis nach einer genaueren
Kartengrundlage gezeigt, die nicht nur quantitative Vergleiche mit den weiter
zurückliegenden Gletscherständen erlaubt, sondern auch als Vergleichsbasis für
die jährlich beobachteten kleineren Veränderungen dienen kann. Das Entgegen-
kommen der Eidg. Landestopographie in Bern und das Geophysikalische Jahr haben
die Bahn freigelegt für die Neuaufnahme 1957 im Masstab 1:10'000, von der als
erstes Blatt 3 soeben erschienen ist und diesem Band der Procès-Verbaux beiliegt.

Das gesamte zu zwei Drittel vergletschte Einzugsgebiet umfasst rund 200 km²
und liegt auf etwa 46½° N und 8° E. Kulminationspunkt ist das Aletschhorn mit
195 m ü.M. Der Hauptstrom legt vom Jungfraujoch (3475 m ü.M.) bis zum Glet-
schertor (ca. 1490 m ü.M.) eine Horizontalstrecke von 22.3 km zurück. Die Firnlinie
verläuft in den letzten 20 Jahren zwischen 3230 m ü.M. und 2760 m ü.M., bei einer
mittleren Lage auf etwa 2900 m ü.M.. Das Verhältnis von Nährgebiet zu Zehrgebiet
schwankt je nach Firnlinienlage von 0.6:1 bis 3:1 bei einem mittleren Wert von
und 1.9:1.

1. BEMERKUNGEN ZU DEN AUFNAHMEARBEITEN

Die Photoflüge wurden in einer Höhe von 5200 m ü.M., mit vollautomatischer
Wildkamera RC 5 mit Objektiv Aviogon mit Brennweite 11.5 cm und Filmformat
8/18 cm durchgeführt. Ausser den Passpunkten auf festem Gelände wurden auch
schwarze quadratische Platten von 1 m², durch schwarze Russringe von ca. 3 m
Durchmesser ergänzt, als Kontrollpunkte auf dem Gletscher verlegt. Wiederholte
Einmessung erlaubte, Lage und Höhe dieser beweglichen Punkte für den Zeitpunkt
des Fluges zu bestimmen.

Die Tatsache, dass es nicht möglich ist, im Stereoautographen die Höhenlinien
für vollständig weisse, kontrastfreie Firnflächen zu ziehen, zwang uns, Kontraste
künstlich zu schaffen. Zu diesem Zweck wurde die Firnoberfläche durch schwarze
Flecken von je ca. 1½-3m Durchmesser gefärbt (für eine Flughöhe von maximal
etwa 2000 m über Terrain). Pro km² waren 300-400 Punkte notwendig. Pro Punkt
brauchten wir ca. 200 gr Material, das sich zu 93 Gewichtsprozenten aus sehr
feinkörnigem, trockenem Sägemehl und zu 7% aus Industrieruss zusammensetzte.
Wir haben versucht, durch systematische Anordnung der Punkte Geländeformen wie
Kuppen und Mulden hervorzuheben. Die Auswertung im Autographen hat aber

gezeigt, dass eine zufällige Verteilung vorzuziehen ist. Das Streuen des Sägemehl-Russ-Gemisches geschah teils von Hand durch Skipatrouillen, teils durch Abwurf aus Flugzeugen, wobei zwei verschiedene Systeme von «Bomben» verwendet wurden. Die erste Art bestand aus Plastic-Packungen, in die Petarden mit Zeitzündschnur eingebaut waren, die zweite aus Packungen in Kartonschachteln, welche sich beim Aufschlag öffneten und dabei durch eine vorgespannte Feder überstülpt wurden. Total wurden 30 km² Fläche mit etwa 10'000 Punkten markiert, wobei ca. 6300 Punkte von Hand, 3500 mit Federschachteln und 200 mit Petarden vom Flugzeug aus gefärbt wurden. Die Arbeit von Hand erfolgte von vorbereiteten Depots aus mit 12 kg Säcken, wobei pro km² mit einem Aufwand von 2-4 Arbeitstagen zu rechnen war.

Dank der Verschmutzung sollte der mittlere Höhenfehler der Aufnahme auch im Firngebiet den in m gemessenen Wert von $\pm (1 + 3 \text{ tg } \alpha)$ nicht überschreiten, wobei α die Hangneigung bedeutet. Im Ablationsgebiet dürfte der mittlere Fehler innerhalb von $\pm (0.5 + 3 \text{ tg } \alpha)$ liegen.

3. KARTOGRAPHISCHE GESTALTUNG

Ursprünglich war ein reines Kurvenbild mit Eintragung nur der Spalten und Gletschergrenzen beabsichtigt. Im Laufe der Autographenauswertung wurde aber der Wunsch nach einer eigentlichen glazial-morphologischen Karte lebendig. Schliesslich erfolgten Auswertung und kartographische Gestaltung nach folgenden Gesichtspunkten:

- a) Das Kartenbild umfasst das gesamte Einzugsgebiet.
- b) Der mittlere Höhenfehler soll innerhalb der unter Kapitel 2 erwähnten Beträge von $1 + 3 \text{ tg } \alpha$, resp. $0.5 + 3 \text{ tg } \alpha$ bleiben.
- c) Der Masstab von 1:10'000 soll für ein möglichst vollständiges Bild der Oberflächengestalt ausgenützt werden.
- d) Grundsätzlich sollen möglichst alle Einzelheiten lagegetreu wiedergegeben werden.
- e) Die Darstellung hat so zu erfolgen, dass alle Einzelheiten gedeutet werden können.
- f) Die Relieferung soll nicht nur eine Betonung des Kurvenbildes sein, sondern zusätzlich Eigenheiten der Morphologie der Oberfläche zeigen, die im Kurvenbild allein nicht zum Ausdruck kommen.
- g) Nach Möglichkeit sollen alte Gletscherstände mitkartiert werden.
- h) Das Randgebiet wird weniger detailliert reliefiert als der Gletscher; auf Felszeichnung wird verzichtet.

Einzelheiten wie Spalten, Strudellöcher, Oberflächenbäche, Strukturlinien, grössere Felsblöcke, Begrenzung der Moränen, Toteisausscheidungen, alte Gletschergrenzen, wurden mittels des Autographen lagerichtig eingezeichnet. Die Farbe charakterisiert das Material. Strukturlinien sind in der Materialfarbe dargestellt in der sie offensichtlich werden. In Toteisgebieten sind die Höhenkurven blau gezeichnet. Die Grenzen zwischen dem nicht überall mit Sicherheit feststellbaren Toteis und dem eigentlichen «bewegten» Gletscher sind durch eine blaue gestrichelte Linie angedeutet.

Besondere Aufmerksamkeit wurde der Wiedergabe kleiner isolierter Firnflächen im Randgebiet geschenkt, deren weitere Entwicklung ein sehr empfindliches Indiz für die kurzfristigen Klimaschwankungen sein wird (vgl. Karte: Firnflecken am Eggshorn).

An der Relieferung wurde vom Kartographen nicht nur im Büro nach Kurvenbild und Luftbildern, sondern wiederholt auch im Gelände gearbeitet. Nur so war es



Bild 1 — Blick vom Silbersand (vgl. Karte 1:10'000) auf den Grossen Aletschgletscher. Aufnahme F. Müller, Mitte Oktober 1957.



Bild 2 — Grosser Aletschgletscher : Luftbild des Abschnittes Märjelen-Zungenende. Aufnahme der Eidg. Landestopographie, 7. September 1957.

möglich, Einzelheiten, wie sie Bild 1 zeigt, im Rahmen der durch den Masstab begrenzten Möglichkeiten darzustellen.

Man war sich von vornherein bewusst, dass der grosse Karteninhalt bei einfarbiger Reproduktion nicht in erwünschtem Masse lesbar sein konnte. Auch Versuche mit schwach aufgedruckten farbigen Flächentönen ergaben kein durchwegs klares Bild.

Nur durch das Mittel des Farbauszuges wurde eine eindeutige Wiedergabe möglich. Vergleichen wir das Luftbild (Bild 1) mit der Karte, so sind wohl gewisse Einzelheiten wie der helle Randstreifen, der Mitte des 19. Jahrhunderts noch vom Eise bedeckt war, im Luftbild ausserordentlich deutlich. Einzelheiten aber, wie sie die komplizierte Oberflächengestalt bei der Einmündung des Mittelaletschgletschers aufweist, können erst durch die kartographische Gestaltung eindeutig erklärt werden. Nur blaue Schraffen beispielsweise lassen blanke Eisböschungen erkennen.

4. VORLÄUFIGE RESULTATE

Von der Aletschkarte 1:10'000, welche den Stand vom Herbst 1957 wiedergibt, ist erst das beiliegende Blatt 3 vollendet. Erst in diesen Tagen wird die Autographenarbeit für alle 4 Blätter, d.h., für das gesamte Einzugsgebiet des Grossen Aletschgletschers fertig. Unter diesen Umständen ist es nicht möglich, bereits einen vollständigen volumetrischen Vergleich mit früheren Gletscherständen zu geben. Trotzdem sollen die Bilder 3 und 4 mit einigen vorläufigen Resultaten über die Veränderungen mit der Zeit informieren.

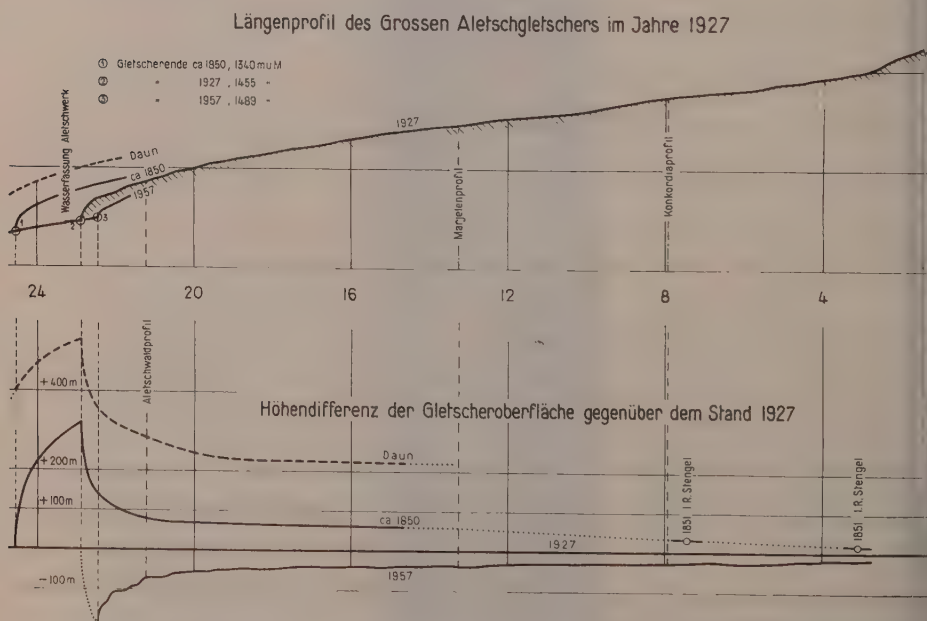


Bild 3 — Längenprofil des Grossen Aletschgletschers und Höhenänderungen der Gletscheroberfläche.

In Bild 3 sind das Längenprofil des Jahres 1927 und die Höhenänderungen der Oberfläche bis 1957 aufgezeichnet. Ausserdem wird der Stand 1927 demjenigen von

ca. 1850, sowie dem Daunstadium [9] gegenübergestellt. Bild 4 enthält die Höhenänderungen 1926/27-1957 in Abhängigkeit der Meereshöhe und die aus dieser Beziehung und der hypsographischen Fläche berechneten Volumänderungen. Die mittlere Höhenänderung in diesen 31 Jahren beträgt rund $16\frac{1}{2}$ m oder 0.5 m/Jahr.

Interessant ist auch der Vergleich mit dem Hochstand des vergangenen Jahrhunderts. J. R. Stengel hat 1851 mehrere Oberflächenpunkte in der Nähe der Firnlinie kotiert. Tabelle 1 zeigt, dass der Gletscher in der Gegend der Firnlinie von 1851-1957 um 0.33-0.56 m/Jahr dünner geworden ist. Ueber die Ursachen dieses Schwundes wurde schon andernorts berichtet [7].

TABELLE 1

Höhenänderung der Oberfläche des Grossen Aletschgletschers von 1851 bis 1957 in 2 Punkten, oberhalb und unterhalb der mittleren Firnlinienlage (ca. 2900 m ü.M).

a) Oberflächenkoten in m über Meer:

Punkt	Aufnahme Stengel Herbst 1851	terrestrische Photogram- metrie Herbst 1926	Absteckung durch Vor- wärtseinschnitt Herbst 1947	Luftphoto- grammetrisch Herbst 1957
III	2993.0 ⁽¹⁾	2978.3	2963.4	2957.9
VIII	2789.0 ⁽¹⁾	2757.2	2737.1	2729.1

b) Kotendifferenzen in m/Jahr:

Punkt	1851-1926 (75 Jahre)	1926-1947 (21 Jahre)	1947-1957 (10 Jahre)	1851-1957 (106 Jahre)
III	- 0.15	- 0.71	- 0.55	- 0.30
VIII	- 0.38	- 0.96	- 0.80	- 0.53

5. SCHLUSSBEMERKUNGEN

Die in Ausarbeitung begriffene Karte 1957 des Aletschgletschers erfordert einen aussergewöhnlich grossen Aufwand. Es ist klar, dass eine periodische Nachführung in dieser Art nicht möglich ist. Doch sind alle Elemente für eine periodische Beobachtung des Gletschers, wie z.B. das Kurvenbild, darin enthalten. Darüber hinaus gibt diese Karte eine detaillierte Bestandesaufnahme in einem genau bekannten Zeitpunkt; schon der Vergleich der hypsographischen Kurven der verschiedenen selbständigen Gletscher des Einzugsgebietes dürfte zu interessanten Schlussfolgerungen führen [1, 2]. Glazial-hydrologische Fragen, wie Einfluss von Exposition

(1) Die Höhenangaben in der Aufnahme Stengel sind auf den alten Horizont bezogen. Deshalb müssen die für das Jahr 1851 in der Tabelle eingetragenen Höhen um 3.2 m herabgesetzt werden, damit sie mit den Zahlen für die Jahre 1926, 1947 und 1957 vergleichbar werden.

Volumänderung des Aletschgletschers von 1926/27 — 1957

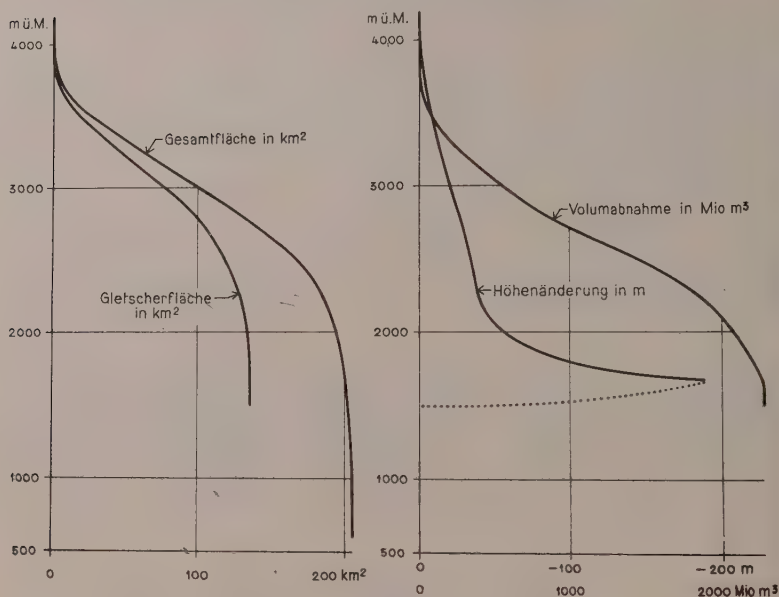


Bild 4 — Volumänderung des Grossen Aletschgletschers 1926/27—1957.

und Lokalklima auf das Regime, lassen sich auf dieser Basis behandeln. Solche theoretische Studien werden wiederum die Grundlage für zielbewusste Felduntersuchungen über die Zusammenhänge zwischen meteorologischen Grössen, Ernährung und Auflösung der Gletscher, Finliniervariation und klimatische Schneegrenze liefern. Die getreuen Bilder von Spalten und Strukturen dürften auch zu Studien anregen, welche sich mit Problemen der Gletschermechanik und der Kristallographie befassen. Ganz abgesehen von der rein kartographischen Bedeutung dürfte diese Spezialkarte vor allem als Planungsgrundlage für weitere wissenschaftliche Untersuchungen aller Art wertvoll sein. Schon wurde z.B. von verschiedenen Seiten angeregt, die neue Aletschkarte für geologische und morphologische Studien zu benützen.

Es ist nicht möglich, im Rahmen dieses Berichtes die vielen Mitarbeiter zu nennen. Allein für die Feldarbeiten waren gleichzeitig rund 40 Vermessungsspezialisten, Piloten und Alpinisten eingesetzt.

Besonderer Dank gebührt der Direktion der Eidg. Landestopographie in Bern, welche die Aufnahmearbeiten geplant, bei den Feldarbeiten mitgewirkt und für die Autographenauswertung, die kartographische Gestaltung und die Reproduktion eine Reihe ihrer besten Spezialisten eingesetzt und damit den weitaus grössten Teil der gesamten Arbeit übernommen hat, ferner der Direktion der Versuchsanstalt für Wasserbau und Erdbau an der Eidg. Technischen Hochschule in Zürich, deren Abteilung für Hydrologie und Glaziologie die Hauptlast der Hauptlast der Feldarbeiten getragen und bei der glazialmorphologischen Gestaltung massgeblich mitgewirkt hat. Neben diesen beiden Institutionen ist die Gletscherkommission als Mitinitiant zu erwähnen. Der Schweiz. Nationalfonds zur Förderung der wissenschaftlichen Forschung hat durch seinen im Rahmen des Geophysikalischen Jahres gewährten Beitrag die Arbeit unterstützt. Schliesslich wurden die Feldarbeiten durch die Gastfreundschaft der Hochalpinen Forschungsstation Jungfraujoch erleichtert.

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GLACIOLOGICAL RESEARCH OF THE ITALIAN KARAKORUM EXPEDITION 1953-1955

ARDITO DESIO — ANTONIO MARUSSI — MICHELE CAPUTO

The glaciological research carried out by the Italian Karakorum Expedition has a two-fold character. On the one hand was the collection of descriptive data on the majority of the glaciers existing within the territory visited by the Expedition, and on the other, several specific detailed glaciological investigations on such of the glaciers as deserved attention because of some particular reason, and which lay within the field of study of the Expedition.

The first was an extensive investigation, while the second, for various reasons was concentrated on only two glaciers: the Baltoro, one of the largest in the Karakorum Range; and the Kuthiah, much smaller in size, but of particular interest because of the exceptional advance of its front foot (12 km) verified in 1953.

We shall refer first to the extensive investigations, limiting ourselves however solely to noting the situation of the fronts of some glaciers and their more recent variations.

Baltoro Glacier — Insofar as the general aspect of the front of this glacier in various photographs put together by Professor Desio, and which are dated 1909, 1929, 1954 and 1955, 'does not appear appreciably different (*), some details and some measurements made by him can supply some precise data on relatively recent variations of the front. It should be remembered that the painted marks which he set up in 1929 were not found again in 1953 and 1954, while a photogrammetric survey of the front is still waiting to be plotted before a comparison can be made with the 1929 survey.

The most positive evidence at present available is that of a great rock standing at the front of the glacier which figures in the photographs taken by V. Sella in the summer of 1909. However, De Filippi, who described the front of the glacier (*), did not leave a note of the distance of the rock from the front. The rock was observed in 1913 by Dainelli (³) who judged its distance as 80 m from the front, and by Desio in 1929 who found it in contact with the front. That these observations dealt with the same rock is proved by the photographic records. The rock is in fact of grey gneiss and is cut diagonally by a white quartz vein which makes it readily identifiable, as well as its characteristic shape. The same rock was found again by Desio in 1953 and its distance from the front was then 296 m (**). In the course of 24 years the front of the Baltoro appears to have retreated about 300 m. The same rock is visible in a photograph kindly made available to Professor Desio by Dr. F. Maraini (1958), but a measure of the distance is lacking. A comparison of the photographs shows no appreciable difference to the eye.

In summing up we can only say that the glacier had an advancing phase between 1909 and 1929, and a retreating phase between 1929 and 1953, but we do not know when these phases began.

The retreating phase of the front of the Baltoro between 1929 and 1953 had parallels also in several lateral glaciers, and among these in the first place the Liligo Glacier. In 1929 the terminal tongue of the glacier, near the left side of the Baltoro, appeared to be rather more wasted than that which figured in a photograph of V. Sella in 1909.

(*) For the variations before 1909 see: Desio A., *La Spedizione Geografica Italiana al Karakorum*, 1929, Milano 1936, p. 393 e segg.

(**) Found by stadia measurement.

In 1953 the front of the Liligo Glacier was separated by about 750 m from the Baltoro.

It seems then that in 1929 the progressive stage indicated by the front of the Baltoro may have been at an end, and that there may have already begun a retrogressive stage which was already making itself felt in the lesser glaciers, such as the Liligo.

It should be remembered that other lateral glaciers from 1929 to 1953 have shown a notable retreat. Thus for example the small Llungka Glacier, to the right of the Baltoro, had in 1929 stopped at several hundred metres from the right side of the larger glacier, in 1954 its front had become independant and was separated by about 1 km from the right side of the Baltoro.

Other variations might be deduced from a comparison between the topographic maps drawn up at different times, but this would require a very close critical examination of the fronts which is hardly appropriate here.

Biafo Glacier — Rather more than the Baltoro, the Biafo Glacier has shown clearly recent variations of its front, recognisable also (in a qualitative way) from photographs. This is due to its peculiar topographic position. As is known, the front of the Biafo spreads out in front of the mouth of a lateral valley of the Biafo River valley, stopping at a certain distance from the cliffs opposite the mouth, where runs the river, which is fed principally by water from the melting of glaciers which lie further up stream, among which are the Baltoro and the Panmah.

In 1929 a reconnaissance topographic survey was made of the environs of the front, but as the survey made in 1954 has not yet been plotted, no comparison is possible. In the occasion of Professor Desio's first visit 5 painted targets were set up, which have not since been found again. Nevertheless a sufficiently certain reference datum exists. In 1929 during the reconnaissance survey it was noted that the most prominent point of the glacier was found on the left side at a distance of about 180 m from the left-hand cliffs of the Biafo valley.

The observations made by Professor Desio in 1953 enable one to estimate the distance of the front from the same cliff as about 1 km. But the ice in that year was so covered by moraine that it was rather difficult to establish the true end of the glacier. However the above mentioned distance should be considered as an approximate minimum, from which the retreat between 1929 and 1953 could not have been less than 820 m; almost three times that of the Baltoro.

Kuthiah Glacier — This glacier in 1953 was subjected to an extraordinary increase so much that in the shortest time there was formed a ice stream 12 km long, to which Professor Desio has previously had occasion to refer ⁽⁵⁾. This phenomenon which he considered in the summer of 1953 prompted him to have photogrammetric survey made of the whole glacier during the 1954 trip by the expedition's surveyor, Major Lombardi. The survey was executed on a scale of 1:50,000. Furthermore a survey was made on the same scale of the near-by glaciers of Goropha and Ranga, which previously had not figured at all on the topographic maps (fig. 1). The Goropha glacier has a length of about 9 km and an area of 7 km².

We do not consider ourselves bound to deal further on this occasion with the causes of the exceptional advance of the Kuthiah. We can in any case note that the other glaciers of the region at the same time showed signs of regression.

Hispar Glacier — Professor Desio visited the front of the Hispar in the summer of 1954 and had the impression that it showed signs of recent wasting. But as that was his first visit, he did not have available a basis for comparison. On the basis of the topographic map on a scale of 1:253,440 of Shipton's Expedition of 1939, he did

not note any appreciable differences in the position of the front, from which it would not seem that the glacier has undergone any notable variations in the last 15 years.

Barpu Glacier — The front of this glacier was also visited in 1954 by Professor Desio. Its chaotic aspect did not allow the observation of signs of retreat, nor of advance. Compared with the topographic map of 1939, there did not appear in 1954 to be any appreciable differences.

1. INVESTIGATIONS BY TOPOGRAPHIC AND GEOPHYSICAL METHODS

Glaciological investigations by topographic and geophysical methods of the Italian Karakorum K2 Expedition of 1954 had the Kuthiah and Baltoro glaciers as their object.

In this communication we shall give an account of the results, not hitherto published, which deal with the photogrammetric and geophysical work carried out in the course of the campaign Expedition of 1954.

The use of photogrammetric methods had a double aim: both of obtaining the topographic features relief of the glacial valley and of the body of the glacier itself, and also of determining the superficial flow velocity of the glacier.

The use of geophysical methods on the other hand had above all the aim of determining the thickness of the ice and the shape of the bed of glacier; which was obtained by two completely distinct methods.

In one method, the theory of which is due to C. Somigliana, the thickness of the ice is determined, assuming a cylindrical glacier of semielliptical cross-section, from the knowledge of the distribution of surface velocities, of the slope of the glacier, and of the coefficient of viscosity of the ice, as well as its density.

In the other method, however, the depth and the shape of the cross-section of the bed are determined, still assuming a cylindrical glacier, from the distribution of gravity anomalies measured across the particular section.

1.1. *Photogrammetric surveys*

These were made by Major Francesco Lombardi of the Istituto Geografico Militare, using a 13×18 cm Zeiss camera with three lenses objectives of focal length of about 19 cm. The surveys were supported in part by being referred to trigonometric points existing in the region, and in part by means of astronomical stations, always measuring directly the photogrammetric base, and where necessary carrying out small local triangulations.

The fundamental heights were determined by means of hypsometry, or were deduced from those of the surrounding trigonometric points.

In particular, for the two surveys of the glacial valleys, the following may be noted:

1.1.1. *Kuthiah Glacier*

This glacier occupies the mountainous portion of Stak Valley, and has a fall from the WNW to the ESE, arising directly from the massive spire of Mt. Haramosh (7397 m). Flowing into it are the Nong and Kosomber glaciers on the left bank, and the Choskiange on the right bank. The foot of the glacier is situated at a point with coordinates $\varphi = 35^{\circ}45'$, $\lambda = 75^{\circ}03'$ where the Stak Valley bends sharply towards the SW.

The survey of the glacier was made by Major Lombardi and was based on two astronomical stations of which one was in the Shoman locality and the other near

the Choskiango glacier, as well as on the following trigonometric points of the Survey of Pakistan.

- no. 121 — Pk. 57/43 I (Thanmari)
- » 122 — Pk. 58/43 I (Haramosh)
- » 123 — Pk. 59/43 I (Korang Kar)
- » 124 — Pk. 60/43 I (Shinka Mashkila)
- » 1 — Pk. 1/43 M (Paraber)

To carry out the survey a total of 19 stations were set up, of which 15 were used; the plotting was completed by the Istituto Geografico Militare on a scale of 1:40,000; the map, which will be published on a scale of 1:50,000, shows, besides the Kuthiah glacier, also the minor glaciers of Goropha and Ranga.

We may note further that Professor Marussi has made a series of observations of the velocity of the Kuthiah glacier in the vicinity of Kulankae, at about 1 km from the front foot of the glacier itself. The observations were made during the days 17, 18, 26 30 June and 4 and 18 July 1954, by means of azimuthal measurements with a Wild T2 theodolite set up at Kulankae (2992 m). The various points observed were identified by targets, or by natural features. The distances of the objects observed were determined by means of a stereoscopic rangefinder.

The maximum velocity encountered on the glacier amounted to 21.9 cm per day, and the distribution of velocities is shown in figure 1.

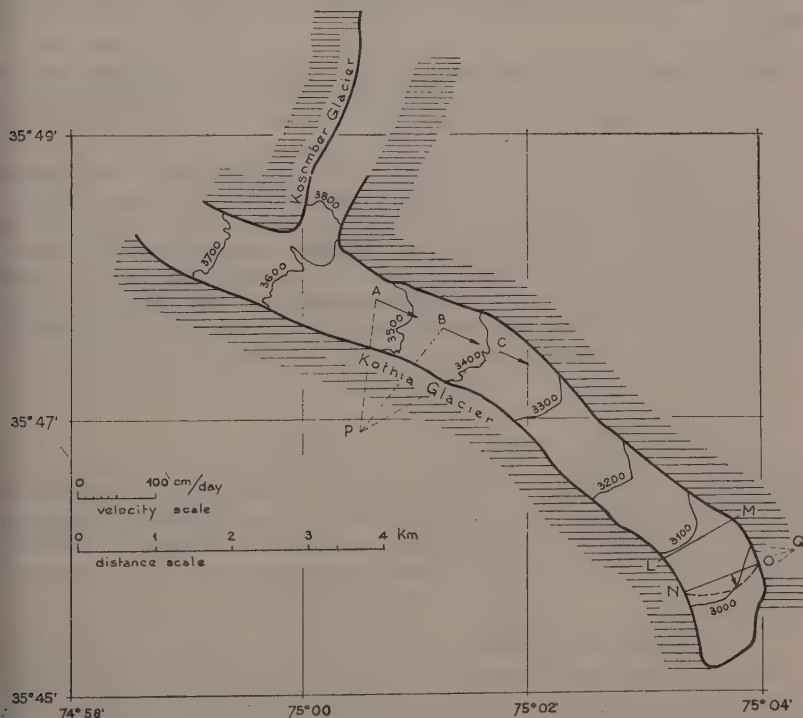


Fig. 1 — Kuthiah Glacier — P & Q = Theodolite observation points
 LM = Gravimetric section
 ON = Section showing distribution of surface velocity.

Professor Marussi further carried out, from the elevation marked 4150 m to the East of the Choskiango, observations of the velocity of the Kuthiah Glacier at three points situated on the main stream, of which the first (A) was located at about 1 km below the confluence of the Kosomber Glacier, the second (B) at about 1 km further downstream from the first, and the third (C) at yet another kilometre further downstream. The observations were carried out during the days of 10 August 1954 and 16 August 1954 by means of a Wild T2 theodolite; the distances required to deduce the displacements from the parallaxes, and the resultant velocity along the course of the glacier, were obtained afterwards, from the photogrammetric map.

The velocities obtained were:

59.2 cm per day for the point A
 54.0 cm per day for the point B
 38.2 cm per day for the point C

1.1.2. *Baltoro Glacier*

It is well known that the Baltoro Glacier was photogrammetrically surveyed as long ago as 1909, as a task of the Expedition of the Duke of the Abruzzi; the results appeared as a map on a scale of 1:100,000. This map was greatly improved during the 1929 Expedition of the Duke of Spoleto. The result of this work appeared as a map on a scale of 1:75,000 which served as the basis for the new surveys of Major Lombardi.

These consisted of a total of 17 stations, which included the highest peak of K2, the northern slopes of Broad Peak, the southern slopes of Broad Peak, and the Conway Saddle, Hidden Peak, and in part, the Gasherbrum Group.

The Pakistani surveyor Badshah Jan has since carried out a plane-table survey of the Liligo glacier, which flows into the Baltoro from the left.

Major Lombardi has further obtained from a photogrammetric base situated on the slopes which overlook Urdukass, two sets of three pairs of stereograms, one with axis normal to the base, the other two with axes inclined at 30° to the right and left respectively, which between them cover a large area tract of the Baltoro glacier, from one side to the other. The first set was made on August 17, and a second on September 12 1954, with a view to showing up the deformations to which the surfaces of the glacier had been subjected during the period, and hence obtaining the distribution of surface velocities.

This distribution is shown in fig. 2.

1.2. *Geophysical research*

This dealt with, insofar as concerns glaciology, the determination of the thickness of the glacier and the determination of the profile of its bed. The investigations were conducted as mentioned, in two different ways: that based on the formulas of C. Somigliana, and that which instead sets out from a knowledge of the gravity anomalies across a section of the glacier. We shall examine separately the two procedures.

1.2.1. *Use of Somigliana's Formula*

Professor Carlo Somigliana has shown that it may be possible, at least in theory, to determine the bottom profile of a glacier (assumed to be cylindrical) by means of the distribution of velocities on the surface. When the maximum velocity is to be found at midstream, and if we make an approximation limited to the second order, the bottom profile is reduced to a semi-ellipse. Reducing the general result to this particular case, Professor Somigliana obtains the expression

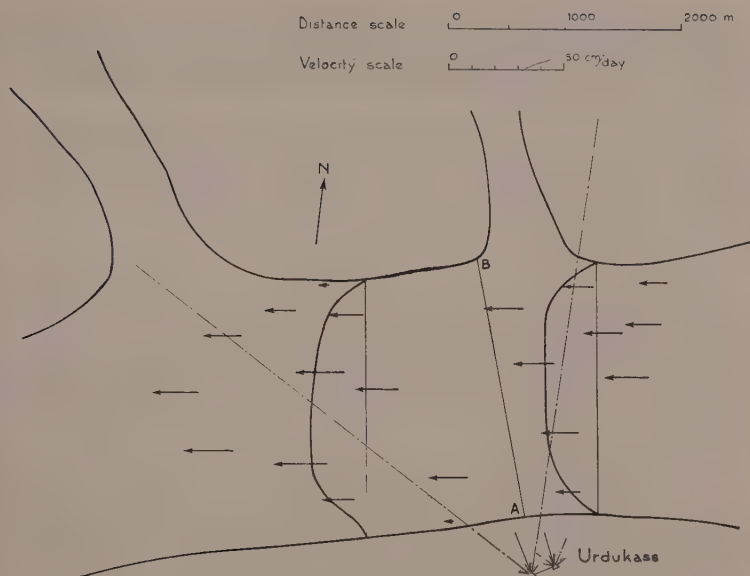


Fig. 2 — Baltoro Glacier at Urdukass—Photogrammetric stations and velocity of the glacier.

$$b = \sqrt{\frac{2\mu L^2(v_0 - U)}{\delta L g \sin \alpha - 2\mu(v_0 - U)}}$$

where b is the minor semidiameter of the ellipse (and hence the depth of the glacier at the middle), L is the major diameter of the ellipse (and hence the width of the glacier), g the gravity, α the slope of the bed, v_0 the velocity at midstream (assumed to be the maximum), U the velocity of the ice at the bottom (zero as a rough approximation), δ the density of the ice, and finally μ the coefficient of viscosity.

	L metres	$\sin \alpha$	v_0 cm/day	b metres	Δg mgal	b' metres
<i>Kuthiah Glacier</i>						
A	800	0.08	59.2	180		
point B	960	0.08	54.0	165		
C	900	0.08	38.2	153		
at 2 km from the foot	1000				10	125
at 1 km from the foot	1000	0.1	21.9	88		
<i>Baltoro Glacier</i>						
Base camp					12	222
Circo Concordia					30	358
Urdukass	2230	0.006	24.7	377	21	390

This formula has been applied by Professor M. Caputo to various cases already examined on the Kuthiah and Baltoro glaciers, where the superficial velocity of flow had been determined. For the density δ the value 0.91 gr. cm^{-3} was adopted and for the coefficient μ the value $135.10^{11} \text{ gr cm}^{-1} \text{ sec}^{-1}$, while the velocity of flow at the bottom is assumed to be zero.

The table above shows the remaining values which are concerned in the various cases, together with the results obtained for the maximum depths. The two final columns, which will be dealt with below, show the maximum gravity anomaly Δg found on the glacier, and the maximum depth b' obtained by the gravimetric method:

1.2.2. Gravimetric method

The gravimetric method worked out for determining the depth of the glacier consists of gravimeter measurements across transverse sections of the glacier, of the terrain and Bouguer reductions of the values obtained, and thus of the determination of the relative anomalies. After this, the anomaly curve, is interpreted on the working hypothesis that the glacier has a cylindrical shape. This leads to a problem of investigation in only two dimensions, which can be solved by methods of approximation.

Such a method of graphical approximation has been elaborated by M. Caputo: it consists of deducing, by a rapid graphical procedure, the anomalies due to an assumed bottom section; comparing the values obtained at various points with those obtained by observation; and making trial modifications to the bottom section until a sufficiently close agreement is reached with the observed values.

Another procedure, more formalized but quicker, also developed by Professor Caputo, consists of assuming that the bottom profile has a semi-elliptic shape. There remains only one parameter to be determined in such a case, the minor semiaxis of the ellipse.

Both procedures have been applied: on the Kuthiah glacier, only that which presumes a semielliptical bottom profile, taken on a section running from Chambot to Kuthiah Hut (3250 m), i.e. about 2 km upstream from the front foot of the glacier. The maximum depth obtained was 125 m, in good agreement with the values obtained by the use of Somigliana's formula, as may be seen from the table given above.

On the Baltoro Glacier, both methods were applied to the gravimetric profiles observed at the Godwin Austen Glacier, at the level of the Expedition's Base Camp, and at Urdukass (fig. 3). The method of the elliptical sections gave results:

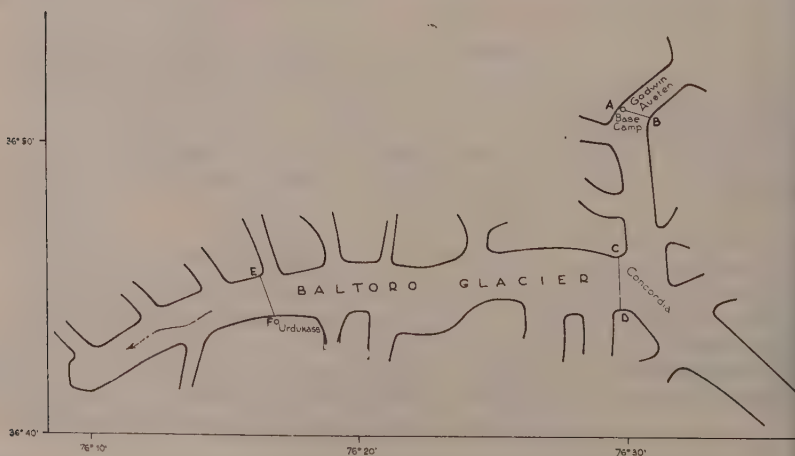


Fig. 3 — Baltoro Glacier sections at which the depth of ice has been determined.

which are shown in the already-mentioned table; while the method proposed for tracing the bottom profile by successive approximations has furnished the results shown in the graphs 4 and 5. In these graphs the curves of the observed gravity anomalies are shown in comparison with the calculated curves.

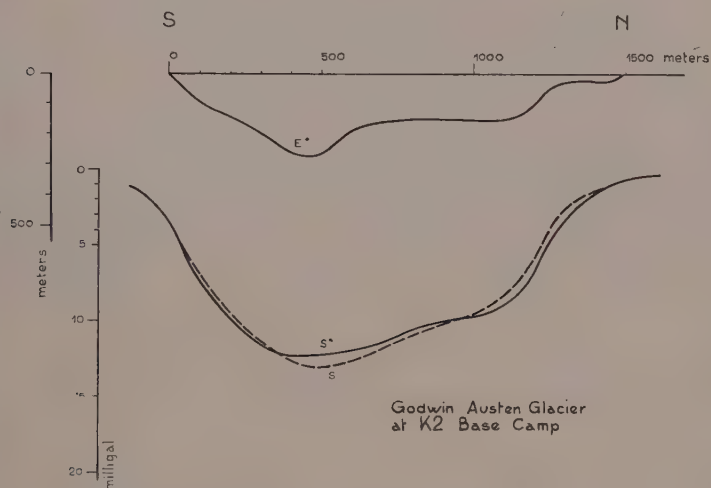


Fig. 4 — E^* = Bottom profile from gravity anomalies
 S^* = Gravity anomaly corresponding to E^*
 S = Observed gravity anomaly.

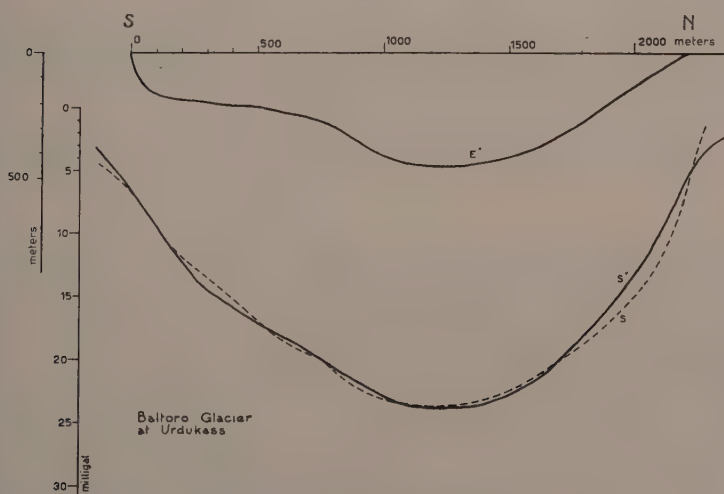


Fig. 5 — E^* = Bottom profile from gravity anomalies
 S^* = Gravity anomaly corresponding to E^*
 S = Observed gravity anomaly.

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PRINCIPAL FEATURES OF THE GLACIERS OF THE NORTHERN SLOPE OF THE DZHUNGAR ALATAU MOUNTAINS

P.A. CHERCASOV (U.S.S.R.)

(Scientific collaborator of the Geographical Department of the
Academy of Science of the Kazakh S.S.R.)

SUMMARY

The Dzhungar Alatau mountains extend South-East of the Kazakh Republic between 44° to 46° N and 78° to 82° E. They are largely represented by two latitude-orientated ridges divided by a deep intermountain depression.

During the 1956-1959 I. G. Y. and I. G. C. period the area investigations cover the glaciers of the Northern slope of the North end of the Dzhungar Alatau mountains within the basin of the Lepsa, Terekty, Baskan and Sarkan Rivers which constitute its centre.

The paper deals with a brief summary on a number of aspects of the glaciological investigations such as :

- 1) General features of the extension of the present glaciers; their quantity, total area, types and their ratio to the total amount and area.

- 2) The paper deals with the characteristics of the principal features of glaciation : extension of the present glaciers due to the absolute height of their position, connection of the glaciation with the upper boundary of the névé fields; the position of the névé line due to the glacier exposure, elements of glaciation and hypsometric data.

- 3) The dynamics of valley glaciers is given by elements : intersecular dynamics of the glaciers and their state within the I. G. Y. period during the recent intersecular climatic cycle; movement of the glaciers, dynamics of the glacier ice, their space state, thickness and volume.

- 4) Hydrological regime of the glaciers consists of two stages : 1 - thawing, - glaciers flow and its share in the feeding of the rivers.

While defining the quality of the hydrophysical processes occurring on the surface of the glaciers the problem of radiation and thermal balance was solved by P. Kuzmin's method for the ablation periods during a number of years with the different point of cloudiness.

The amount of the surface thawing of the glaciers made it possible to calculate for a number of years with different current of cloudiness :

1. the thawing of ice and snow due to the temperature of the air,
2. the total thawing of ice and snow due to the height of the place for the ablation period.

The latter was the ground for the calculation of the glacier flow and its share in the total flow of the rivers at the source as well as at the outlet on to the foothill valley where their waters are used in national economy.

RÉSUMÉ

Les monts de l'Ala Tau de Djoungar sont situés dans le Sud-Est de la R.S.S. de Kazakhie entre 44° et 46° de lat. N et 78° et 82° de long. E. Ils présentent essentiellement deux chaînes orientées suivant la latitude divisées par une profonde faille tectonique.

Durant la période de l'A.G.I. (1956 à 1959) les recherches furent entreprises sur les glaciers de la face Nord de la chaîne septentrionale de l'Ala Tau de Djoungar dans les limites des bassins des fleuves : Lepsa, Terekty, Baskan et Sarkan qui y occupent une position centrale.

Le rapport donne sommairement les conclusions essentielles sur plusieurs parties des recherches glaciologiques :

- 1) particularités générales sur l'emplacement des glaciers contemporains : leur nombre, leur surface totale, leur type;

- 2) caractéristiques des caractères essentiels de la glaciation : propagation des glaciers contemporains en fonction de la hauteur absolue; rapport de la glaciation avec la limite supérieure des champs de névé; disposition des névés en fonction de l'exposition des glaciers; éléments de glaciation et d'hypsométrie;

3) la dynamique des glaciers des vallées par éléments : dynamique interséculaire des glaciers et leur condition durant la période de l'A.G.I. dans le dernier cycle climatique interséculaire, la progression des glaciers, la dynamique de la substance glaciaire, la condition spéciale, la puissance et le volume des glaciers;

4) le régime hydrologique des glaciers comprend deux parties :

1 — fonte et 2 — écoulement des glaciers et son rôle dans l'alimentation des fleuves.

Au cours de la détermination de l'aspect qualitatif des processus hydrophysiques se déroulant sur la surface des glaciers (pour les périodes d'ablation durant plusieurs années avec points de nébulosité différents) le problème du bilan thermique et de la radiation de la surface des glaciers, suivant la méthode de P. Kouzmine, fut résolu.

L'aspect quantitatif de la fonte superficielle des glaciers permit de calculer pour plusieurs années avec nébulosités différentes : 1) — la fonte de la glace et de la neige dues à la température de l'air et 2) — la fonte totale de la glace et de la neige due à l'altitude du site durant toute la période de l'ablation.

Ce dernier rapport servit de base pour les calculs de l'écoulement des glaciers et sa participation dans l'écoulement général des fleuves aussi bien aux sources qu'à sa sortie dans la plaine située au pied des monts où ses eaux sont largement mis à profit par l'économie nationale.

1. The mountains of the Djungar Alatau are situated in south-east Kazakh S.S.R. between the 44° and 46° north latitude and 78° and 82° east longitude. They almost 400 km in the latitudinal direction. The longitudinal valleys of the rivers Koksui, in the west and Borotala in the east, divide the Djungar Alatau into two large ridges: the Northern and the Southern. The northern ridge, elevated to 4.000 m and higher over the sea level, has in the central part, on the North slope a region of a wide development of contemporary glaciation (fig. 1).

The present report intends to give a brief characterization of the conditions of existence of contemporary glaciation of the range of Djungar Alatau, an example of which is given by the basins of the rivers Lepsa, Terekty, Baskan and Sarkan, that occupy in the Djungar Alatau a central position, during the period of the years from 1956 to 1959.

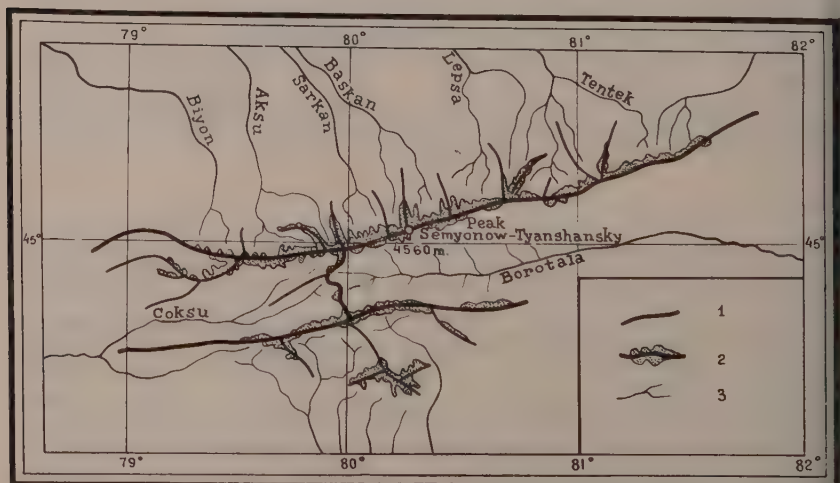


Fig. 1 — Orographical scheme of the Djungar Alatau.
1) ridges and spurs, 2) glaciers, 3) rivers.

2. GENERAL SITUATION PECULIARITIES OF CONTEMPORARY GLACIERS

Within the area of the mountainous part of the basins described, the glaciers occupy 9%.

They can be divided into two large groups: 1-glaciers of valleys that are situated between the slopes of separate ridges. The largest glaciers of the valley type reach with their open tongues a height of 2.950-3.000 m. Parts of the tongue that are burried under the «cloaks» of the frontal moraines go down below the open ends another 50-300 m.

The distribution of the different types of glaciers according to their number and area is shown in the table 1.

TABLE 1

No in rotation	Groups and types of glaciers	Quantity of glaciers	entire area Km ²	Proportion to quantity of all glaciers of basins	Proportion to area of all glaciers of basins %
1	<i>I. Valley glaciers</i>				
2	of valleys	39	147,34	20	57
3	of cavities	1	21,23	0,5	8
4	of hanging valleys	28	34,38	16	13
	<i>II. Glaciers of Mountainous slopes</i>				
5	Hanging	83	28,34	43,5	11
6	Trailing	8	7,53	4	3
7	of cirques	25	15,70	13	6
	of flat summits	6	2,01	3	2
Total		190	256,53	100,0	100

The largest of all glaciers of the ridge Djungar Alatau known at present is the cavity glacier Djambul. Its entire area extends to 21,23 km².

A number of valley glaciers of minor dimensions have the following areas: Berg—16,68 Km², Aba—13,21 Km², Satpaev—9,67 Km², Gerasimov—8,00 Km².

3. THE PRINCIPAL FEATURES OF CONTEMPORARY GLACIATION

Owing to general climatic conditions that maintain the existence of contemporary glaciation, hypsometry plays a major role in the distribution of large masses of ice among separate parts of the axis section of the ridge, in conjunction with ice is determined by orography and mechanical redistribution of snow under the influence of prevailing winds.

For glaciers exposed to the north, the following connection of the glaciation area with the upper edge of the firn fields has been ascertained; $P = 0.0029 (B - 3550)$.

Here P = the area of the glacier (with the buried parts), in km^2 that affects 1 km of the upper edge of its firn fields; B = the average altitude of this edge measured in m; the number 3.550 = the lower limit of the altitude, measured in m, at which the value $P = 0$; the number 0,0029 = the gradient of the value P , i.e. the area of glaciation that affects 1 m of vertical extension of firn field of the glacier.

Basing upon this connection, it has been ascertained: 1—that as a lower limit, at which glaciers can be formed and exist on the slopes of the north exposition, it is the isohypsos of 3.550 m; 2—by the permanently acting factor of formation of glaciers that equals to 0,0029, for highly elevated regions of the ridge, the average altitude of which reaches 4.000-4.560 m, for each 1 km of extension of the upper limit of firn fields corresponds to 1 km^2 of ice, whereas for the medium altitude of mountain crests of the lower regions of the ridge (not higher than 3.600-3.900 m),—only averagely about 0,45 km^2 .

The firn line being dependent upon the exposition, is situated on the glaciers of the north exposition at an average altitude of 3.550 m, and on the glaciers of the south exposition—at the altitude of 3.900 m.

The glaciers of the central part of the ridge occupy only 67% of their receptacle. Their general area decompose into separate elements as follows: firn fields—39%, open tongues—38%, side and central moraine—80% and buried parts of the tongue—18%. At the whole, of the described basins the positive difference of glaciation (280-300 m) surpasses the negative one by (225-230 m).

The average glacial coefficient in 1956 was 0,80-0,81, but in 1957, owing to the lowering of the snow line—, it was 1,30-1,34. The prevailing area of the contemporary glaciation (about 85%) is concentrated within the altitudes of 3.300 to 4.100 m. In this interval the largest area of glaciation falls upon the altitudes from 3.500 to 3.700 m, whereby in the more elevated regions of the ridge the main areas of glaciation are concentrated 160 m lower than in the still lower regions (Fig. 2).

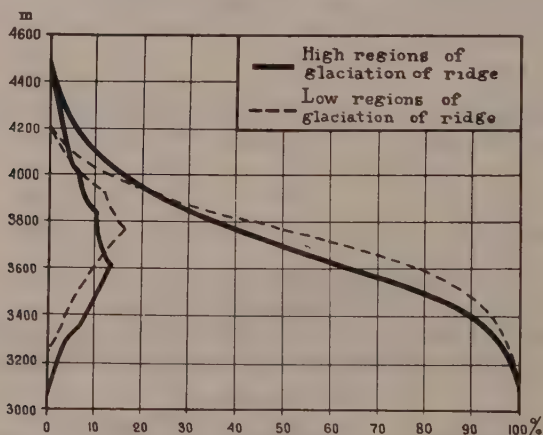


Fig. 2 — Comparison graphic of hypsographical curves of distribution and hypsometrical curves of glaciation area by separate altitudinal regions.

4. DYNAMICS OF THE GLACIERS

For characterization of the course of main events that are taking place in the glaciers during the period of the I.G.Y., it was necessary to find the spatial and

temporal place of this observation period within the general period of recession of the existing glaciers since the time of their utmost development.

It has been ascertained that the contemporary glaciers have evidenced a form of oscillation, at an average duration of 35 years of each period. Up to the years of 1957-1959, since the time of their maximum development, five intracentennial oscillations have elapsed. The last two or three oscillations that occurred in a number of ridges of Kazakhstan and Middle Asia (including the Djungar Alatau) since the end of XIX century until the middle of the XX century, were registered by means of direct observations. These oscillations represented; in 1878-1880—the advance of glaciers; 1900-1903—an intensive recession of glaciers; 1915-1918—an advancing or a stationary condition; 1930-1935—an recession of glaciers; 1948-1950—slackening of recession, and by some glaciers—a stationary condition was observed. Consequently, in the years of 1963-1965 on the glaciers of Kazakhstan and of Middle Asia must enter a moment of an utmost active reduction of glaciers.

The cyclicity of recession of the glaciers in the form of rhythmically repeating oscillations at intervals of about 35 years is quite closely connected with similar cyclicity of level changes of outletless lakes, ascertained by A. V. Snitnikov (1957) because the life activity of both glaciers and lakes is entirely dependent on one and the same factor—the climate.

From the above it, follows that the processes which are taking place in the glaciers of the Djungar Alatau (which will be spoken of below) during one period of the International Geophysical Year, happen at a time that occupies, approximately, an intermediary position between the extreme points of a 35-years-climatic cycle in the life of glaciers, that has begun at 1948-1950.

The dynamics of glaciers are being analyzed by the examples of the largest valley glaciers: Berg, Satpaev, Shtshukin, Gerasimov, Abay, Shumsky and Djambul. The following issues were here considered: 1) Movement of glaciers; 2) Dynamics of glacial substance; 3) Spatial condition; 4) thickness of glaciers; 5) Volume of glaciers.

1. The movement of the majority of glaciers is conditioned chiefly by the incline of the bed. Depending on the arrangement of the receptacle, the movement is fulfilled according to the «opposed extrusion flow» type or the «opposed gravity flow» type (Avsyuk, 1948). The motion speed of glaciers with «opposed-gravity flow» is as a rule, higher, because the accumulated potential energy of motion is released here in a considerably narrower space of diametral (transversal) profile than by glaciers of the former types.

The average yearly motion speed of glaciers moving by the «opposed extrusion flow» type, amounts to 1,10-1,20 m on the lower part of the stream and 3,60-3,80 m on its upper part.

By glaciers with «opposed gravity» type of motion, the speeds amount to 4,0-4,3 m on the lower part, and on the upper part of the stream (near the firn line)—from 10,5 to 12,0 m in a year, whereby the average daily motion speed during the summer months are 1,3-1,4 times higher than those of the average daily all-year speeds.

2. The yearly balance of mass in the district of the glacier tongues shows that the efflux of ice mass elevates the glacier surface, on the average through the entire tongue area, up to 30 cm, and melting lowers the same surface at an average of 110 cm. Hence the yearly impariment of the thickness of the glacial tongues equals, approximately, a layer of 80 cm.

The loss of ice caused by melting is compensated by the efflux, that equals to 30% of the value of melted masses during one year's period, and about 10%—during the ablation period.

The mentioned percentage proportions show that the tongues of the glaciers receive from their firn fields quite insufficient feeding. Furthermore, approaching the end of the tongues, the values of the efflux diminish more and more.

During a year's period the ice surface on the firn line elevates 13-14 cm; the line where the decrease of ice begins, is placed lower. On this line the value of the year's ablation that equals, on the average to about 20 cm, is balanced by an equal amount of a year's efflux. This balancing line is situated at altitudes of 3.400-3.500 m, in 300-500 m from the firn line. Such distance amounts to 11-12% of the entire length of the open tongues. Hence the conclusion is that 88-89% of the length of the tongues belong to those parts that are subject to diminishment.

3. The decrease of thickness of the glaciers influences their spatial condition. Depending on conditions of environment, feeding and motion, in the mentioned years of observations, the glaciers lost from 0,005 km² to 0,027 km² of their area, that in proportion to the area of the tongues, it amounts to from 0,2 to 0,6%. In linear measurements this decrease amounted to, according to the above mentioned condition, from 6,5 to 30 m.

4. The thickness of the glaciers was determined by method of balances (N.N. Palgow, 1958) and Lagally formula (Kalesnik, 1939). It has been ascertained that depending on the arrangement of the glacial receptacle, the distribution of the thickness of the glaciers is similarly different. Glaciers of the cavity and semicavity type, have, as a rule, more thickness of ice near the terminus of the glacier than at the firn line. Near the terminus the thickness reaches 100-130 m, but near the firn line 80-100 m.

In this case is observed a specific example of «damming up» of the ice masses that causes an increase of the glacier's thickness of given type by their sliding down the valley.

For glaciers having classical forms of receptacles, the thickness gradually diminishes toward the terminus. In the region of the firn line, the thickness reaches 100-150 m, but near the ends of the tongues — 40 — 70 m.

The exactness of indirect methods of ascertaining of the thickness of glaciers is confirmed by seismographical sounding.

5. It was ascertained that the decisive importance in the distribution of glacier ice masses through action of contemporary climatic conditions, belongs to the orography of the basin and to hypsometry.

By glaciers with less favourable conditions of existence, the volumetric glacier coefficient is 0,50, and by better conditions — 1,40.

The general quantity of mass lost during the mentioned years of observations amounts for a number of glaciers in proportion to their volume, only to 0,3-0,4% which will ensure their long existence.

As it is shown by the material above, the contemporary glaciation of the Djungar Alatau is gradually reducing in its dimensions.

V. The hydrological regime of glaciers includes two sections: 1 — thaw, and 2 — the flow of the glaciers and the part it is playing in feeding the rivers.

1. The quantitative side of glacier melting under conditions of contemporary climate was determined by boring of bars into the ice.

For determination of the qualitative side of the hydrophysical process that occur on the surface of glaciers during the ablation period of 1950-1957 (which had a different course of meteorological conditions), on the glaciers of the basin of the river Baskan were organized actinametrical and gradiental observations and upon their basis was solved the problem of the thermobalance of glaciers by method of P.P. Kusmin (1948).

It has been determined that in the zone of contemporary glaciation of the ridge of the Djungar Alatau, the main influence on the melting of the glaciers (provided, their majority enjoy almost similar orographical repository) is rendered by solar radiation. On the second place there is the action of the local thermoexchange with the air which is, however, often complicated by advective heat.

The analysis of observations showed, that the significance of the solar radiation in the melting of glaciers directly depends on the value of the radiation balance ($W\odot$). The latter, however, in its turn, is considerably influenced by cloudiness.

In 1956, at an average state of cloudiness during the entire ablation period of 4 points, the radiation balance reached during 24 hours 273 cal/cm^2 , but in 1957, at a state of cloudiness of 7 points—it was only 107 cal/cm^2 within 24 hours.

The second component of the thermo—balance—the convectional heat, including also the heat arriving as a result of condensation of hydrovapours ($Wk + 600 \text{ F}$), varies not in direct proportion to the variation of the absolute values of convectional thermoexchange, but it depends on the fluctuation of the radiation balance.

Thus, in 1956, at a state of cloudiness of 4 points, the relative contribution of the radiant heat ($W\odot$) that was spent on the surface ablation at an altitude of 3040 m, reached 74%, but the contribution of convectional heat, together with the heat received as a result of condensation of the hydrovapours ($Wk + 600 \text{ F}$), was only 26%. In 1957, at an average state of cloudiness of 7 points during the entire ablation period, the relative contribution of $W\odot$, in comparison with 1956, at the same absolute altitude, went down to 57%, whereby the share of ($Wk + 600 \text{ F}$) increased to 43%, the share of the incoming heat due to condensation of the hydrovapours (600 F) in both observation years was insignificant: 1956 it amounted to 3% and in 1957—to 6% (Fig. 3).

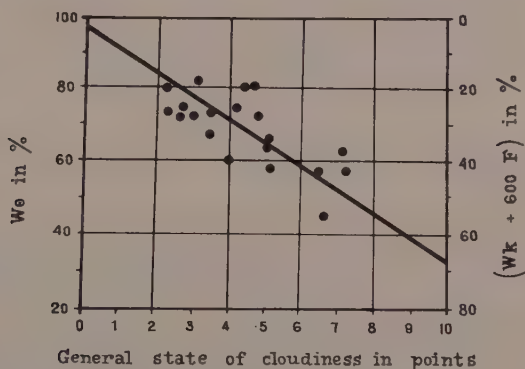


Fig. 3 — The dependency of the relative importance (in %) of components of the thermal balance up on general cloudiness. ($W\odot$)—radiation balance, ($Wk + 600 \text{ F}$)—convectional thermo-exchange and thermoexchange in the system evaporation—condensation.

As it will be seen from the above, the greatest part of the surface (H) ablation is conditioned by processes of melting in which the leading place belongs to the solar radiant energy and the auxiliary place to the thermoenergy that is coming from the air in a direct form. A very small part in the surface ablation belongs to evaporation, which is amply compensated during the summer time by condensation of hydro vapours on the surface of the glaciers.

Thus, on the average for the entire ablation period of 1956, at the altitude of 3040 m, out of the entire amount of the ablation during 24 hours (H) (4,560 cm of ice, converted into water), the effect of ($W\odot$) melted 3,527 cm, the effect of Wk —1,051 cm. At the same time, however, the condensation of hydro vapours (600 F) diminished the value of surface melting (h) (equals 4,578 cm)—by 0,018 cm.

In 1957, accordingly, (H) amounted to 2,200 cm; out of this, $W\odot$ melted 1,399 cm, Wk —0,821 cm—therefore, $h = 2,220 \text{ cm}$ and condensation diminished the value of ice melting by 0,020 cm.

Therefore, in the Djungar Alatau by determination of the superficial ablation by means of boring of bars into the ice, it can be considered without any obviously perceptible deficiency in the exactness of observation, that the melting volume (h) equals the value of the superficial ablation (H).

The calculation of melting of the glaciers in the mountains by data of the thermo balance is quite embarrassing, and are actually possible only at those places where actionometrical observations had been made.

A most elastic meteorological element of climat is the air temperature. On the one hand it very well reflects the balance of the solar radiation and, on the other hand, it occurs as such a meteorological element, that it can be determined comparatively exactly for any unexplored point of this or other part of the ridge.

As a negative feature of establishing a connection between the air temperature and the melting of the ice mass is, that this connection is not universal. In each concrete case, dependent on the course of average cloudiness and air moistness during the observation period, the significance of the air temperature, varies also as a factor comprising the thermobalance. This leads to such a condition that the values of the melted snow and ice, averaged during 24 hours may be different for equal air temperature, although the percentage contents of the snow in the melted masses may be the same. In each case of observation of the melting, is it necessary to establish a separate dependency on the air temperature, that would respond to the course of the ice melting by given cloudiness.

For the year of 1956, for the entire ablation period by a with an average state of cloudiness of 4 points, the value of melted mass (containing 98,6% of ice and 1,4% of snow) was determined by the equation:

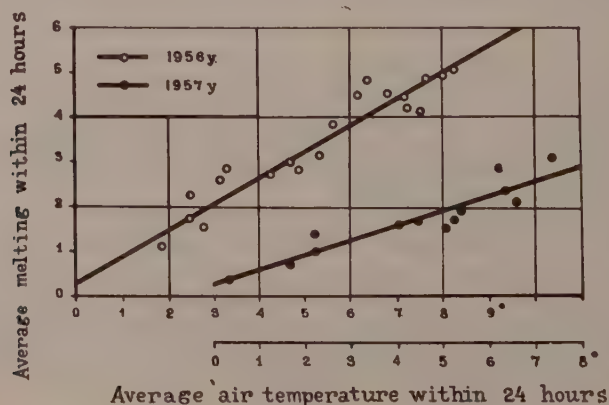
$$A = 0,58. \quad t + 0,30.$$

For the year of 1957, with an average cloudiness of 7 points and similar contents of snow in the melted mass, —by the equation:

$$A = 0,33. \quad t + 0,22.$$

Here: A = the average melting of ice during 24 hours in cm (converted into water); t = the average air temperature during 24 hours at the height of 2 m above the observation points of the melting during the same observation period; the numeral significations-, are the parameters.

In the interval of the average air temperatures during 24 hours from 0° to 10° , the melting due to one degree of warmth was in 1956 6,1 mm, but in 1957-3,5 mm, but in 1957-3,5 mm, with a precision of 8 per cent (Fig. 4).



In connection with the above, the melting values of the mass (converted into water) depending on the altitude of the place, will equally be different during the entire ablation period. For 1956 mentioned dependency has been expressed by the equation.

$$Tl = 0,405 \frac{No - N_1}{100} \text{ and for 1957 } Tl = 0,296 \frac{No - Nl}{100}$$

where Tl —the total ice melting in m during the entire ablation period; no . — the altitude of the firn line; N_1 —the altitude, for which the melting value is being determined (Fig. 5).

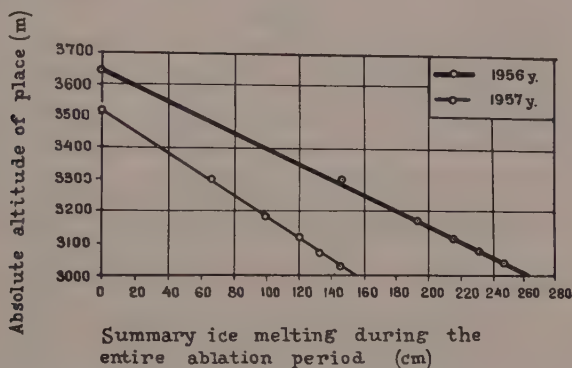


Fig. 5 — Dependence of ice melting up on the altitude of the place during the entire ablation period.

The melting calculations of snow accumulated during the winter on the glacier tongue surface were made depending on the altitude of the place, by formula:

$$\text{for 1956 } T_s = 0,00116 (N - 3000) - 0,030;$$

$$\text{for 1957 } T_s = 0,00084 (N - 3000) - 0,020$$

where T_s —the total melting of the winter snow (converted into water) in meters, during the entire ablation period on the glacier tongue; N —the absolute altitude, for which the melting value is being determined.

In 1956, with a state of cloudiness of 4 points, the melting of snow for 1° of warmth amounted to 5 mm (converted into water), and in 1957, with a state of cloudiness of 7 points—3 mm.

This regularity of distribution of melting values of ice and snow on the glaciers which depends on the altitude of the place was taken as a basis for calculation of flowing of the glaciers.

2. During the time of flowing of the rivers, that arise by snow and ice melting on the glacier zone in the warm part of the year, two periods had been singled out:

1) Flow that arises by melting of snow and ice during the «general ablation period» (i.e. a flowing as a melting result of the entire sum of solid precipitations during the year and of the melting of ice until the end of the ablation period) and

2) Flow that arises by melting of snow and ice during the «particular or glacial ablation period» (i.e. the flow that arose by melting of the snow that remained on the glacier since the moment of freeing of its end from the snowy cover, and by melting, of the ice until the end of the ablation period).

The second period comprises the greater part of the entire melting time.

In the forming of the melting water on the glacier, these areas participate:

1) the ice of the open tongue, 2) the ice that lies under the side and middle morai-

nes, 3) the ice of the buried parts of the glacier tongue, 4) the firn fields, 5) the snow that lies on the area of the glacier basin which is free from the glaciation and also the snow on the open and buried tongue.

During the general ablation period the total flow from the glacier area (from all melting parts mentioned above) in the region of glacier basins, reached in 1956 76%, and in 1957--69%, and the remaining 24% and 31% of the flow are from the precipitation from the bare parts.

The significance of the glacier water in the flow of rivers from the glacier basin area during the glacier ablation period (in comparison with the flow during the entire melting period) increases in 1956 to 94%, and in 1957 to 83%, and while the role of the precipitation diminishes accordingly.

Owing to unequal value of the surface melting, the average yearly flow increment and the flow increment during the glacier ablation period for the mentioned observation years turned out to be different.

The average yearly flow increment out of all sorts of flow from the area of glacier basins, was in 1956 28 l/sec. and in 1957--20 l/sec. During the glacier ablation period the flow reached respectively 151 l/sec. and 67 l/sec.

Such large difference between flow increments for two adjacent years of observation was due to different melting intensity of the glaciers, different quantities of precipitations, etc.

This circumstance leads to irregular feeding of the ridge rivers with melted glacier water.

On the average, the glacier water in the ridge rivers in the source region amounts to 72% of the yearly flow, and at the mountain exit--only 23%. The remaining 77% of the flow was covered by snow-, rain- and underground water, coming from the territory of the basins that is not occupied by glaciers.

During the glacier ablation period, the glacier component of the river flow (in comparison with the year's flow) increases sharply. At the sources it reaches 89%, and at the mountain exit it amounted to 47%, i.e. twice as much than during the year's period.

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SOME THEORETICAL RESULTS OF THE GLACIOLOGICAL EXPLORATION IN ALTAI DURING THE INTERNATIONAL GEOPHYSICAL YEAR

M.V. TRONOV (U.S.S.R.)

SUMMARY

1. Some theoretical problems of the glaciology were worked out on the basis of the exploration of several years in Altai : on principle of the conformity, on dam of the glaciers, on two levels of the border of snow etc. They were printed mainly in the papers « Problems of the mountain glaciology » (1954) and « Problems of the connection between the climate and the glazing » (1957). Now we have some new results on this problem.

2. The observation of the summer snowfalls let draw conclusions on two general glaciological problems : 1) on influence of such snowfalls on the dynamics of glaciers, on their special role in the usual alternation of the various types of weather; 2) on influence of the spreading surface on processes of the snow-accumulation and the snow-melting.

3. Especially significant were the facts of the observation over the great snowfall on the 25-27 of August, 1957, when the thickness of snow amounted to 50-80 cm. in the region of the upper valley and the glaciers Aktru. The actual duration of the period of ablation was shortened approximately by 20 days.

The count of the balance shows that the annual renewal of such snowfalls can cause the advance of glaciers. The snow was on the surface of glacier for 15-17 days longer than on stones, lying on the same altitude. The influence of the spreading surface is equivalent to the change of the altitude approximately for 400 m.

4. The observations in August 1959 on the plane watershed near the boundary of the snow level proved the distinction between the conditions of the formation and the preservation of glaciers. On the background of the considerable shortening of main glaciers this year a considerable growing of the border of a high situated small glacier with new field of ice took place. Its formation is connected with cold summer seasons of the years 1957-1958, and its preservation in 1959 is connected with the change of the spreading surface.

5. The numerical characteristics of the effects of the dam of ice can be given for the Big Aktru glacier in which the reason of the dam is evident and the thickness of ice is determined. The rise of the level of the surface of glacier is 150-175 m, the increase of the length of glacier is 1,5-2 km. The glaciers Right and Left could not join their edges without that dam.

RÉSUMÉ

1. Les investigations de longue durée sur l'Altai servirent de base pour l'élaboration d'une série de problèmes théoriques de glaciologie, comme le sont : le principe de conformité, le « sanglage » des glaciers, les deux niveaux de la limite de neige, etc. Elles sont, principalement, exposées dans les livres : « Principes de glaciologie des montagnes » (1954) et « Principes du rapport entre le climat et la glaciation » (1957). A présent l'on a accumulé sur ces problèmes un matériel nouveau.

2. Les observations sur les *chutes de neige d'été* permettent de tirer des conclusions en rapport à deux questions, portant sur la glaciologie générale : 1. l'influence de telles précipitations sur la dynamique des glaciers, leur rôle particulier dans l'alternation habituelle des types divers de temps; 2. l'influence de la surface sous-jacente sur les processus d'accumulation de neige et de fonte de neige.

3. Particulièrement démonstratives sont les données des observations sur la grande chute de neige, qui eut lieu le 25-27 août 1957, quand la puissance du champ de neige dans la région de la vallée supérieure et des glaciers Aktru ont atteint 50-80 cm. La durée réelle de la période d'ablation diminua approximativement par 20 jours. Les calculs du bilan montrent qu'une répétition annuelle de chutes de neige semblables peut provoquer une transgression des glaciers. La neige se maintint sur la surface froide des langues du glacier de 15 à 17 jours plus longtemps que sur les pierres à altitude égale. L'influence de la surface sous-jacente est égale à une modification de la hauteur absolue de près de 400 m.

4. La différence entre les *conditions de la naissance et de la conservation des glaciers* se manifeste dans les observations, faites en août 1959 sur le partage plane

des eaux près du niveau de la limite de neige. Sur un fond de forte diminution générale des glaciers principaux, il eut lieu cette année-là un accroissement considérable de l'extrémité d'un glacier, disposé à une grande hauteur, par un champ de glace nouveau. Sa formation est liée aux saisons froides d'été en 1957-1958 et sa conservation en 1959 — au changement de la surface sous-jacente.

5. Les caractéristiques numériques de « l'effet de sangle » peuvent être évaluées pour le glacier Bolchoi Aktrou, sous lequel est claire la cause du sangle et définie la puissance de la glace. La hausse du niveau de la surface glaciaire est de 150 à 175 m; l'augmentation de la longueur de la langue du glacier 1,5-2 km. Sans ce sangle, les glaciers Aktrou Droit et Aktrou Gauche ne pourraient point s'unir par leurs extrémités.

Researches in the Altai glaciers carried on for a period of many years have formed the basis for working out a number of theoretical problems of glaciology, such as the principle of conformity, glacier damming, the two different levels of the snow-line, and others. These problems have been set forth in the writer's books "Problems of Mountain Glaciology", (1954), "Problems of Interrelation between Climate and Glaciation", (1956), and "Principles of Glacioclimatology", (microfilm, 1959), as well as in a number of papers published in journals and symposiums. The general ideas underlying all the above works of the writer may be stated as follows.

Glacier changes and fluctuation over both short and long periods of time cannot be regarded as completely dependent on relief and meteorological conditions. The course of glacier evolutions is, to a great extent, determined by the inherent features of its development. Whatever changes there may occur in the size and condition of the glacier, these changes, in their turn, come to influence the course of the further glaciation process. This influence under certain conditions may become a factor of decisive significance and give rise to "self-development" stages of the glacier. One can distinguish several aspects in the big and active role the glacier itself plays in the glaciation process, such as: 1) changing the underlying surface from stone to snow-firn-ice and vice versa; 2) changes in glacier height; 3) changes in glacier surface conditions; 4) breaking and restoring glacier continuity.

The total amount of precipitation is little, if at all, influenced by these factors. There must be some influence on the percentage of solid or condensed precipitates, but this influence can hardly be significant, since at great elevations there is but little precipitation in liquid form in general. But the processes of ablation can be very greatly affected by these factors. "The factor of underlying surface" (i.e. changes of the earth's surface), is of fundamental importance, since it may exert a large scale influence over both long and short time periods, and makes it possible to distinguish: 1) conditions of preservation of already existent glaciers, and 2) conditions under which, they may arise. If a flat watershed situated at a high altitude has a broken-stone surface, the snow falling in summer melts away rapidly, but if this watershed is already covered with old snow, summer snowfalls are apt to accumulate new layers of snow.

Changes in glacier height result in the raising or lowering of the glacier surface, thereby influencing the extent of glacier ablation, since the rate of melting decreases with elevation. On steep mountain ranges the influence of this factor is but small, but in other cases it proves to be of considerable importance and may even determine the sign of the material balance of the glacier. Thus, the damming of the glacier may raise the surface of the glacier some hundreds of meters and weaken the ablation process on large areas, which results in a general increase in size of such glaciers. This "glacier height factor" may also be of paramount importance in those cases when the flat glacier bed lies below the snow-line level and the nourishment zone is formed only due to the lifting up of the glacier cap or dome. This is a very frequent instance of "inertia in glacier preservation", referred to by many writers at the present time, and characteristic of Iceland and the Arctic Zone, including Greenland.

The breaking of glacier continuity is very common in the regression stage of glaciation and results in increased ablation processes in the glacial basin; this accelerates the shrinkage of the glacier. In this case the course of glacier development, predetermined by the climate, may be divided into three parts or stages: 1) general weakening of the compound glacier; 2) its breaking up or destruction; 3) additional shrinkage of the separate glaciers. The latter may preserve larger or smaller size according to the altitude of the slopes. It should be emphasized that glacier destruction in itself and the breaking of glaciers into separate parts is as characteristic of the regression phase of glaciation as is the gradual glacier recession; but it can never be accounted for merely by the direct influence of the climate, for some additional stimuli of the process always arise. (cf. e.g. Mannerfelt, 1945; Kinzl, 1953; Tronov, 1954; Bondarev, 1959).

Glaciological observations in the Altai in 1957-1959 were principally conducted on the Aktru glacier in the North-Chuisk mountains and on the Taldurinsky glacier in the South-Chuisk range. The Aktru glacier has an area of about 10 square kilometers; the altitude of its peaks above sea level is 3600-4100 m, and that of the snow-line—2900-3000 m. The Taldurinsky glacier is twice as large in its area; snow-line is at 3000-3100 m, the peaks rise to 4200 m (see Fig. 1, 2, 5).

A general relief feature of the North-and South-Chuisk mountain ranges is the frequent occurrence of broad and flat watersheds with broken-stone surfaces at the altitude of 3100-3400 m, i.e. 200-400 m above the snow-line of the glacier. A temporary snow-cover is not unfrequently formed, even in midsummer, on such watersheds which are not covered with permanent snow and firn fields. The above glaciers also possess the following interesting features: the surface of the former is considerably raised due to the damming of the ice by a rocky barrier, and the Taldurinsky glacier

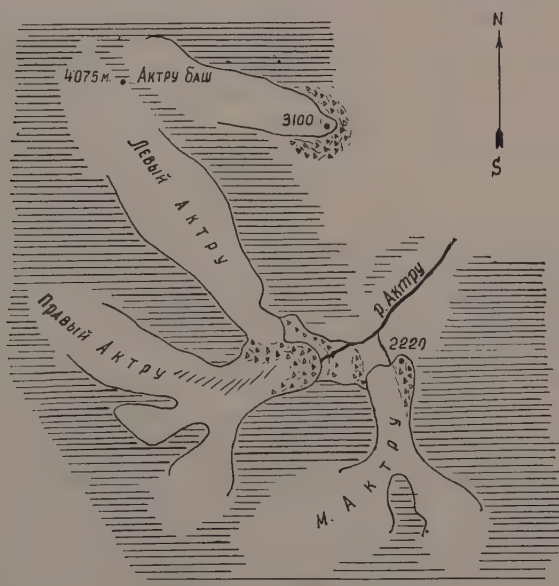


СХЕМА РАСПОЛОЖЕНИЯ ЛЕДНИКОВ АКТРУ

Fig. 1 — Scheme of the Situation of Aktru glaciers.



Fig. 2

appears to have entered the deterioration stage and is shrinking very rapidly at present. These peculiarities of the regions chosen and of the principal objects of research made it possible to obtain some new and interesting data as to the conditions of glacier origination, the factors and significance of glacial development processes which possess features independent of the climate.

Summer snowfall observations in the whole of the Altai mountain region, and not only in the Chuisk district, are conducted under favourable conditions, for snowfall is comparatively frequent; it occurs several times during the summer period. Now and then, unfrequently the low situated termini of large glaciers are covered with new fallen snow. After a heavy snowfall, which generally comes after several days of bad weather, a fine and sunny, though cool, weather usually sets in. The snow on the stones melts away quickly under sun's rays, but it does not do so on the glaciers, and then the white shining surfaces of the glacier tongues are seen with striking distinctness against the dark rocks and moraines. The glaciers glitter reflecting the sun's rays, but do not melt; glacier rivers become shallow and clear. Two problems of general interest in glaciology are consequently connected with summer snowfall observations: 1) the influence exerted by such snowfalls upon glacier regime and the special role they play in the usual alternation of different weather types; 2) the influence of the underlying surface on the processes of accumulation and melting of snow.

The years 1957 and 1958 were rich in summer snowfalls. In the summer of 1957 there were five small snowfalls which, taken together, resulted in a snow cover being formed on the Aktru glacier for a period of 14 days; then another very heavy snowfall occurred in the end of the summer. In 1958 there were two heavy summer snowfalls which formed a snow cover on the glacier tongues for 16 days. One is compelled to draw the general conclusion that under the Altai conditions more or less frequent recurrence of summer snowfalls can be a decisive weather and climatic factor determining the glacier dynamics. This conclusion is confirmed by calculations based on observation data for the heavy snowfall in the end of the summer 1957.

The snowfall on August 25-27, 1957, was an exceptional phenomenon even among heavy snowfalls. The layer of new fallen snow in the Aktru valley near the edge of the forest at the altitude of 2000-2200 m was 40-50 cm thick and in the region of the lower parts of the glaciers and terminal moraines—50-80 cm thick. The snow

cover was rather dense, especially on the glaciers, so that the amount of water in the snow was 20 mm and more. Then fine weather predominated for a lengthy period; but it was cool, and at the beginning of the period even cold, with temperature minima below zero. Up to the first half of September day temperatures of 8-10 prevailed in the wooded parts of the valley, while on the glaciers they did not rise above 5°-6°. Under such temperature conditions solar radiation must be the principal factor of glacier ablation.

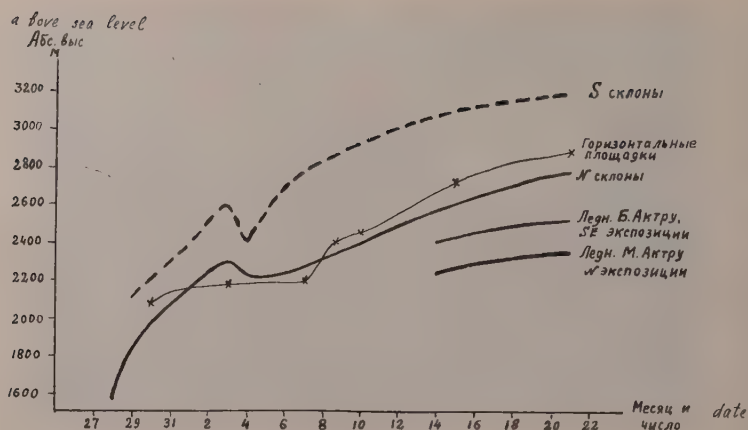
The subsequent melting of the snow was the subject of careful observations up to September 22. In the Aktru valley the snow gradually melted away completely, but not in one or two days as usual, but only by the 6th or 7th of September. By September 21st the snow melted away on the lowest part of the tongue of the Small Aktru glacier reaching down to the altitude of 2220 m northwards; accumulation of ice on the glacier surface also occurred in places. The snow melting line on the tongue of the Big Aktru glacier which had an eastern exposure had by this time moved upward to 2450-2500 m, i.e. not to a very great altitude. On flat watersheds and benches, by September 21st, snow was preserved only at altitudes of 2800-3000 m and above, while the rocky slopes with southern exposure were completely bare of snow.

Fig. 3. Gives a chart representing the course of the snow melting process after the above mentioned snowfall under various conditions. The abscissae are numbers showing consecutive days of observation, whereas the ordinates are the corresponding elevation boundaries of the snow melting line on slopes with southern exposure on moraine platforms and benches and on glaciers. The origin of the curves describing the melting of snow on glaciers is shifted to the right because the baring of glacier tongues began considerably later. Comparing the ordinates of the curves for the same date of observation, one can see that the snow melting line reaches the altitude of 300-400 m on slopes and glaciers with northern exposure, and 700-800 m on glaciers on the southern slopes of the mountain ranges. It is also possible to make up a table giving the number of days with snow cover on glaciers and surrounding locality during the period from August 27 to September 21. Here is the table:

Altitude in m	2150	2200	2300	2400	2500
Aktru valley	4				
Southern slopes	2	2	4	8	8
Northern slopes	5	6	12	14	16
Glaciers with southern exposure				18	21
Glaciers with northern exposure		18	20	25	25

Thus, this snowfall alone accounted for the difference of 16-17 in the number of snow cover days on glaciers and on rocky slopes. The observation data given can serve as material for estimating the influence of the underlying surface factor on the regime of the accumulation and melting of snow. It may be roughly estimated that a glacier surface (in the Altai conditions), as compared with a rocky surface, can increase the annual number of snow cover days approximately by 40 and lower the level of the zero balance of the snow by 400 m. However, further systematic observations are necessary.

There are different possible ways of approach to the task of a quantitative estimation of the influence which is exerted on the glacier regime by summer snowfalls, such as: 1) actinometric measurements and subsequent calculation of losses in solar



Ход границы таяния снежного покрова в Бассейне Антру после снегопада 25-27 августа 1957 г.

Fig. 3 — Migration of the melting line of the snow cover in the Aktru basin after the snowfall on August 25-27, 1957.

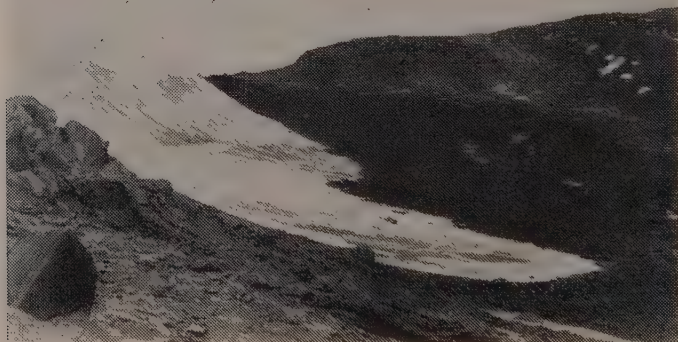


Fig. 4

radiation due to reflexion by the snow; 2) calculation of the actual reduction of the ablation period due to interruptions in the summer melting of glaciers; 3) additional taking into account of heat expenditure on melting the snow layer; 4) hydrometric measurements which point to a poorer water content of glacier streams after snow falls as compared with normal conditions. When the influence of the snowfall referred to above is evaluated by the first and the second of the above ways, the following results are obtained: the shortage in solar heat received by the glacier surface which results from the intensive reflexion of the sun's rays by the snow amounted to approximately 4000 cal per square centimeter, which is equivalent to preservation of an ice layer 56 cm thick on the glaciers; the ablation period was also found to be

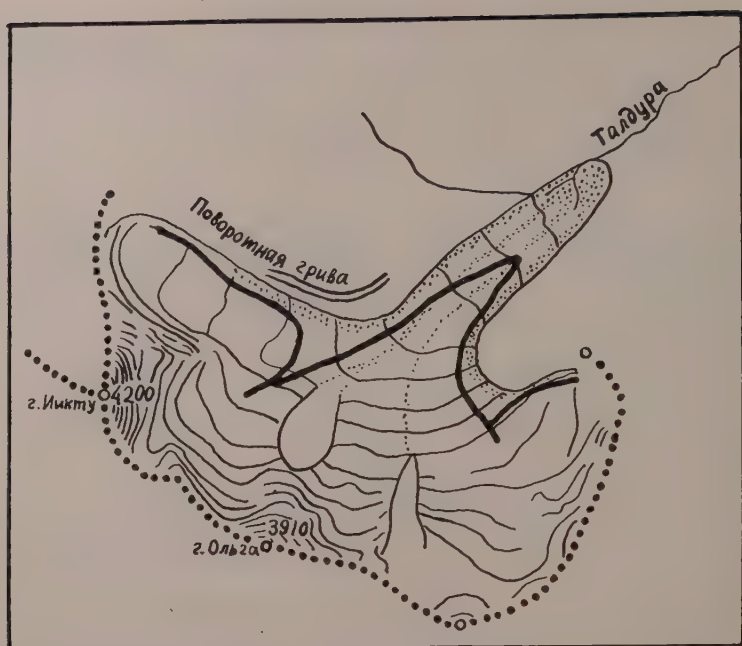
reduced by 18 days, which also corresponds to a decrease of 50-60 cm in seasonal ablation of the Aktru glacier. At any rate, it is reasonable to assume that recurrence of such snowfalls, unless it is compensated by very warm weather periods, might not only stop glacier recession but bring about glacier advance. An analogous point of view as regards the Caucasus was stated by L. A. Vardanyants (1935).

The effect of snowfalls and of the cool summer of 1957 and 1958 could also be detected in the year 1959 which had a warm summer.

The formation of a glacier tongue against the background of overall glacier shrinkage was found in 1959 in one lateral Aktru glacier situated high above the principal glacier to the left of its valley. Observations of fundamental importance was made of some contradictory features in the dynamics of the glacier. This comparatively small glacier, with its terminus situated on a flat watershed at 3100 m covered with broken-stone, i.e. 700 and 900 m above the termini of the Big and Small Aktru, instead of recessing and shrinking was increased (as compared with 1956) by a glacial field 100 m long and wide. It was not an advance of the glacier but it increased in extension all the same. This can by no means be explained by the fact that changes in meteorological regime from 1958 to 1959 might have occurred in a different manner above and below. The main point here is a special process, predicted theoretically, which is, to a great extent, independent of the climate: the snow-ice field once formed here (1957-1958) inevitably acquired the tendency for "self-preservation" due to the reflexion of the solar radiation; as to melting by the direct influence of warm air, it is insignificant at great altitudes. Moreover, the temperatures of the air, which are low at great altitudes, were found to be still more lowered due to the effect of the snow-ice field itself. It should be pointed out that the observation was made in the end of the summer (August 18), when the melting of snow in the Altai mountain ranges drops noticeably. Therefore the new glacial formation described could not shrink considerably before the cold winter spell set in. When the work of the expedition in 1959 was drawing to its close, another case of analogous glacier evolution at a great altitude was discovered, which seemed to contradict the general trend of glacier recession. It is a matter of great interest now to follow the evolution of such elevated glaciers and find out the extent of stability of the newformed snow-ice fields.

The deterioration of the Taldurinsky glacier was described by us in the paper "In the Altai Glaciers" (1951). Fig. 5 represents a scheme showing how several ice streams making up a compound glacier are situated. The general configuration of the large glacier (area of 20 square km) points to inevitable damming of the ice, provided the snow-line is sufficiently low. If the snow-line rises (above a certain boundary), the high level of the glacier surface in the basin can no longer be maintained by annual accumulation of ice, and thus intensive ablation occurs in the greater part of the glacier surface. As a result a process of glacier deterioration develops, which is to a great extent independent of the climate. The theory of glacier damming was dealt with in the book "Problems of Glaciology" (1954)

In the middle of the last century during the general glacier advance, the surface of the Taldurinsky glacier in its central part was approximately 150 m higher as compared with its present level. When the glacier was visited by Prof. V. V. Sapozhnikov, it was retreating but preserved its continuity and had a flat central glacial field. In 1949 not only a rapid decrease of the glacier was recorded (25 m a year and more), but also a sharp weakening of the separate streams and their tendency for separation. There was not a trace left of the central field; in fact, only two streams joined here now which arrived with a considerable gradient and at different levels and preserved different conditions of the surface. The general change in the glacier configuration as seen from the accompanying scheme shows that its deterioration has commenced.



Талдуринский ледник в 1897г. по В.В. Саложникову.
Современный контур ледника дан жирной линией.

Fig. 5 — Schematic chart of the Taldurinsky glacier according to V. V. Sapozhnikov. The present outline of the glacier is drawn in a heavy line.

The powerful factors of the process are the strong erosion of the glacier bulk by water streams, particularly in the ice margins, as well as the intensive melting away of the lateral ice walls. The deterioration of the Taldurinsky glacier resulting from the lowering of its surface may undoubtedly be cited as an example of dammed glaciers evolution in the regression phase of glaciation.

In 1958 G. S. Kravtsov determined the height of the Taldurinsky glacier by the seismic method; the greatest value was found to be 175 m. Deterioration features in the glacier became still more pronounced. Now we must define more precisely the fundamental conceptions as to the essence and sequence of the deterioration process in glaciers of the Taldurinsky glacier type possessing an extensive and comparatively sloping ice accumulation basin and a narrow outlet for the glacier tongue. It should be noted that this type of glacier is very common in many mountain regions.

The height of the glacier (175 m) was not very great and can no longer be the evidence of any considerable ice damming. The accumulation of ice in the basin is mainly compensated by its melting away within the basin as well. Now the narrowest place in the outlet of the glacier tongue lets the reduced in quantities pass freely, without damming. Further shrinkage and deterioration of the glacier is not now caused by the discharge of the basin formerly packed with ice, but it may be explained by the "principle of conformity". Its essence can be briefly stated thus. If a glacier basin is completely filled by the glacier, a conformity is found to exist between climate, relief and glaciation. In the case of an unfavourabl change in the climate,

this conformity is broken; the glacier basin becomes "too large", and then the extensive compound glacier inevitably breaks into several smaller ones. Thus, the entire recession and further deterioration process in glaciers of the type described by us may be caused by a comparatively small climatic change and may be divided into three stages: 1) original recession of the glacier caused by the rise of the snow-line; 2) lowering of the glacier surface level and further shrinkage of the glacier due to the fact that the discharge of ice from the basin is not compensated by its annual accumulation; 3) deterioration of the glacier and its breaking up into parts. Actually these stages may more or less overlap. Still, they must be characterized by various peculiarities in the situation of glacial moraines. This question, however, requires special consideration.

The peculiar features displayed by the Big Aktru glacier (or rather by its left outlet glacier) are also of interest from the theoretical point of view. The upper rather wide valley is blocked from below by a rocky barrier. Here the glacier breaks off into a steep ice-fall, and below, at 2450 m, it forms a small ice-tongue which partly merges with the ice tongue of the second (right) Aktru outlet glacier. Above the barrier, at altitudes above 2700 m, the glacier fills its valley up to the brim and has a sloping flow, although it is furrowed by numerous cracks, sometimes very deep. The glacier has pronounced external features of ice damming, this condition having been preserved since 1936; only the lower tongue retreated a little. The height of the glacier determined by means of the seismic method in 1957 (G.S. Kravtsov) turned out to be very great on the sloping flow part of the glacier; it amounted to as much as 350 m, which can be said to be an unusually great height for a midsized glacier, not exceeding 7 km. The cause is ice damming.

The question may well arise as to the quantitative estimation of the extension increase of this glacier, which is due to damming. No data are available for a precise calculation, but it is possible to make an approximate estimation. For such a glacier as the left Big Aktru stream the "normal" height would be no more than 175 m, and not 350 m. Then the glacial surface level would be also lowered by 175 m, which would result in an approximately 1.4 m increase in seasonal ice melting. The ice volume which now melts away over the 3.5 km extension of the lower glacier flow would melt away over the 2.2-2.5 km extension; the glacier would be over 1 km shorter than it is now. The two glaciers, the Right Aktru and the Left Aktru, would not merge their termini. It may be that this evaluation is somewhat exaggerated as to the lowering of glacier surface, but, on the other hand, it should be taken into account that the glacier would lose several small firn streams on the right and shady slope which at present add substantially to its ice supply.

This example demonstrates the comparatively small but perfectly real influence of damming on glacier extension. It is clear, however, that the process may occur on a much larger scale.

More accurate quantitative data on all the problems referred to above will be obtained in the future. We are of the opinion that both in the Altai and other glacier regions special investigations of the general problem of the inherent features of glacier development should be organized.

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GLAZIOLOGISCHE ARBEITEN DER DEUTSCHEN SPITZBERGEN - EXPEDITION 1959

J. BÜDEL (Würzburg)

SUMMARY

The German Spitsbergen - Expedition 1959 had been acting on Barents - Island (SE-Spitsbergen) from 28th July to 25th August 1959. The main research was done in observing gelisol-activity (congeliturbation and solifluction), the influence by these features to non-glacigenous run off. Observations of the resulting characteristic landforms of the « Frostschuttzone » in comparison with fossil corresponding zones of the ice age in Central Europe also have been made. This expedition has been engaged with preparing the mean-expedition. The latter is planned for 2½ months in sommer 1960. Moreover the expedition observed a series of glaciological questions in connection with gelisol-activity. This paper gives a summary of the results of this research.

Die Expedition war vom 28.7. bis 25.8.1959 auf der Barents-Insel (Südost-Spitzbergen) tätig. Ihr Schwergewicht lag auf der Untersuchung der Frostboden- (Kryoturbations- und Solifluktions-) -Vorgänge, der Beeinflussung der nichtglazigenen Fließtätigkeit durch diese Vorgänge und dem daraus entstehenden charakteristischen Gesamtformenschatz der Frostschuttzone im Vergleich zu der entsprechenden pleistozän-fossilen Zone Mitteleuropas. Sie war damit eine Vorexpedition zu der im Sommer 1960 geplanten 2½-monatigen Hauptexpedition. Das Gesamtunternehmen galt außerdem einer Reihe damit in Zusammenhang stehender glaziologischen Fragen. Hier sind kurz die hierzu erzielten Ergebnisse zusammengefaßt.

1. Die Gesamt-Vereisung Spitzbergens im *Hochglazial* der letzten pleistozänen Kaltzeit (Würmkaltzeit) war nur ein Teil der gewaltigen Inlandeisbedeckung, die damals den größten Teil des Barentssee-Schelfes bedeckte (und isostatisch eindrückte). Die Gesamt-Inlandeisfläche Nordeuropas wird damit um rund 1/3 vermehrt. Indessen hatte das Barentssee-Inlandeis ein *eigenes Vereisungszentrum*, das wahrscheinlich ostwärts des Svalbard-Archipels auf (heute weithin noch untergetauchten) Schelfplateaus im Bereich zwischen König-Karl-Land, Franz-Josef-Land und Nowaja Semlja lag. Wie das Keewatin-Zentrum des pleistozänen nordamerikanischen Inlandeises und das heutige Zentrum des grönländischen lag es also über einer derzeitigen Einmuldung der festen Erdrinde. Diese Ergebnisse wurden u.a. durch eine neuartige Analyse mehrerer Gletscherschliff-Generationen erzielt; sie stimmen mit den — auf ganz anderen Wegen gewonnenen — einer gleichzeitigen französischen Expedition unter I. Corbel gut überein.

2. Im Spätglazial der letzten Kaltzeit wurde das Zurückweichen dieses Inlandeises von seinem Hoch-stand im Bereich Spitzbergens von zwei charakteristischen Phasen verringerter Inlandeisbildung abgelöst. Diese sind aber nicht einfach nur Rückzugsstadien des hochglazialen Inlandeises, sondern hierbei eigenständig aufgebaute, speziell dem Relief Spitzbergens angepaßte Neubildungen entsprechend zweifach länger dauernden — Perioden zunehmender Wärme im Spätglazial. Auch diese Stadien wurden durch die Trennung der zugehörigen Gletscherschliffgenerationen ermittelt. Während der älteren war das Hochland Nordwest-Spitzbergens das beherrschende Haupt-Vereisungszentrum, daneben bestanden nur wenige randliche Zentren von geringer Selbständigkeit. Die Eisbedeckung war bereits auf den Raum Spitzbergens beschränkt, griff aber allseits noch beträchtlich über dessen Außenküsten hinaus

Im jüngeren Stadium dürfte das Inlandeis die Küsten Spitzbergens nicht mehr weit überschritten haben, im Inneren was es — bei im ganzen noch geschlossener, aber wohl dünnerer Decke — bereits viel stärker in lokale Vereisungszentren aufgegliedert. Die Schneegrenze muß demnach noch weitgehend im Meeresniveau gelegen haben.

3. Die heutige Vergletscherung im Hauptteil Spitzbergens mit der gänzlich isolierten Eisbildung seiner einzelnen Gebirgsgruppen und größeren Nebenseln steht auch gegen diesen jüngeren spätglazialen Stand (außer im Nordost-Land) abermals weit zurück : Sie ist wiederum als eine weitgehende Neubildung nach dem wahrscheinlich sehr starkem Gletscherschwund der postglazialen Wärmezeit (ungefähr 4000 bis 500 vor Christus nach Feyling-Hanssen) zu betrachten. Während der jüngsten relativen Warmphase des europäischen Nordpolargebietes (rd. 1920-1940 nach Ahlmann) verlief die Schneegrenze im Bereich des Svalbard-Archipels von rund 600 m Seehöhe im SW bis auf rd. 50 Seehöhe im NO, erhob sich aber in den trockensten inneren Teilen des Eis-Fjordes und der Wijde-Bay bis etwa 900 m. Sie dürfte heute nur wenig tiefer liegen (siehe Abschnitt 6).

4. Die heutigen Gletscher der Barents-Inseln (Südost-Spitzbergens) haben auf die jüngsten kleinphasigen Klimaschwankungen : die Warmephase von rund 1920-1940 und die seitdem wieder erkennbare Abkühlung in sehr eigentümlich verschiedenartiger Weise reagiert, obwohl sie alle von einem gemeinsamen Firnfeld im Inneren der (annähernd quadratischen, 40×40 km großen) Insel ausstrahlen und mit Zungen von durchaus ähnlicher Größe den Meeresspiegel im N, W, S und O der Insel eben berühren. Der westwärts ziehende Duckwitzgletscher reichte 1936 fast 2 km weiter seewärts als heute, 1944 blieb er dagegen rund 1 km hinter der heutigen Mittellage zurück. Der nach S ziehende Freeman-Gletscher war dagegen gerade 1936 sehr viel schwächlicher und rund $2\frac{1}{2}$ km kürzer als heute : er blieb ganz innerhalb der Küstenlinie und zeigte 1944 noch den gleichen Tiefstand. In der Mitte der fünfziger Jahre stieß er aber dann plötzlich sogar noch einige hundert Meter über seinen heutigen Extremstand hinaus, weit über die Küstenlinie vor. In ähnlicher Weise zeigen die ost- und nordwärtziehenden Zungenausläufer des ganzen zentralen Firnfeldes trotz ähnlicher Größe sehr unterschiedliche Vorstoß- und Rückschmelz-Phasen — sogar sich unmittelbar berührende.

Die Erklärung für dieses Verhalten liegt offenbar darin begründet, daß hier kalt-subpolare Gletscher von stark plastisch-druckweisen (nicht viscos-stetig fließenden) Bewegungstypus vorliegen. Sie liegen (insbesondere nach Liestöls Untersuchungen in W-Spitzbergen) oft jahrzehntelang fast bewegungslos, um dann nach Ueberschreitung eines bestimmten Schwellenwertes der Firnanhäufung im Meergebiet plötzlich rasch vorzustoßen. Obwohl Warmphasen diese Anhäufung verzögern, Kaltphasen sie beschleunigen müssen, werden Klimaschwankungen geringen Ausschlages und kurzer Dauer (von äußerstenfalls wenigen Jahrzehnten wie die 11-bis 35-jährigen Perioden) von solchen Gletschern (im Gegensatz zu ähnlich großen temperierten Gletschern, z.B. etwa denen der Alpen) im ganzen nicht synchron abgebildet. Nur Klimawellen größeren Ausmaßes spiegeln sich auch in den Schwankungen kalt-subpolaren Gletscher gleichartig wider (s.o. Abschnitt 1) und 2.).

5. Es taucht die Frage auf, ob durch Klimaphasen mit stärkerer Erwärmung Gletscher von kalt-subpolarem Typus von ihrem Zungenende her in solche vom temperierten Typus umgewandelt werden können. Die Folge wäre ein anderer Bewegungstypus dieser Gletscher und eine raschere und engere Anpassung ihrer Schwankungen an diejenigen des Klimas (auch an kleinere von nur wenigen Jahrzehnten Dauer). Es ist darüber hinaus die Frage zu prüfen, wieweit nun ein- und derselbe Gletscher in seinem Firnfeld dem kalt-subpolaren, an seinem Zungenende

aber bereits dem warm-temperierten Typus verkörpern kann. Fälle solcher Art sind noch nicht untersucht worden. Ihr Vorkommen darf am ehesten in den klimatischen Randzonen des Polargebietes und außerdem dort vermutet werden, wo nicht sehr mächtige Gletscherzungen schwimmend weit in ein (relativ warmes). Meer vorgeschoben werden. Die Ueberprüfung dieser Möglichkeiten ist auf der Hauptexpedition vorgesehen.

6. Die jüngste Abkühlung des Klimas (nach Ahlmann seit etwa 1940) kommt auf der Barents-Insel im Auftreten zahlreicher übersommernder Firnflecke im unvergletscherten Gelände (zusammen 53% der Insel; das vergletscherte Areal umfaßt nur 45% ihrer Fläche) zum Ausdruck. Die meisten dieser Firnflächen waren 1936 noch nicht vorhanden. Ihre oft deutliche Firnschichtung zeigt an, daß sie bereits mehrere Jahre (z. Teil bis nahe an 20 Jahre) alt sind. Andererseits sind sie zum weit überwiegenden Teil noch nicht mächtig genug, um in selbständige Fließbewegung überzugehen und sich in den Tiefenlinien des Geländes zu sammeln : sie kleben meist als «Wächten» noch einseitig an Schneewind- Leehängen. Immerhin sind sie als — vermutlich vorübergehende — Vorboten einer erheblichen Herabdrückung der Schneegrenze aufzufassen, die mit ihren übrigen zäher verschiebbaren Kriterien dann eintreten würde, wenn die derzeitige Abkühlung des Klimas im europäischen Nordpolargebiet (gegenüber der jüngsten Warmphase zwischen 1920 und 1940) länger andauern sollte. Für die Beurteilung der glaziologischen Verhältnisse Spitzbergens und deren temporärer Wandlung im Zusammenhang mit längeren und kürzeren, eng — und weiträumigen Klimaschwankungen ist daher außer der laufenden wissenschaftlichen Beobachtung einiger ausgewählter größerer Gletscher auch diejenige der kleinen Firnfläche, sowie — s.o. Abschnitt 1-3 — die systematische Verfolgung der fossilen Gletscherspuren aller Art notwendig.

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GLACIER DOMES WITH FIRN ALIMENTATION ON THE FRANZ-JOSEPH LAND

A.N. KRENKE

SUMMARY

Glacier domes are one of the most important elements of the glacier sheets on the Franz-Joseph land. The majority of them enters into the alimentation region of the firn or ice type. (The ice type of alimentation corresponds to the alimentation of glaciers, belonging to the Baffin type). The aim of the report is a clearing up of the differences between the alimentation types of the domes and a characteristic of the vital activity of domes with firn alimentation. As an example of the latter is chosen the Jackson dome on the Hooker isle, which is compared to the dome Churlianis on the same isle with alimentation of the ice type.

According to data of snowmeasuring surveys, the quantity of solid deposits on the Jackson dome is approximately the same as on the Churlianis one; it equals 300-350 mm in the water layer. On the other hand, the great height, in comparison to the Churlianis dome, and its remoteness from the sea, leads to less favorable ablation conditions. The difference in ablation value explains the difference between the alimentation types on the summits of both domes.

The temperature and radiation regimes of the periphery parts of the Jackson dome contribute in a greater measure to ablation, and this, in turn, is reflected upon the types of ice-formation. On the Jackson dome are distinguished three vertical zones: zone of firn alimentation, zone of ice alimentation and the ablation zone. The height of the firn limit is 325-390 m; the one of equilibrium line—270 m over the sealevel.

The firn zone on the eastern slope of the Jackson dome adjoins the firn basin of the Obrutchev glacier, that descends lower than the firn limit of the dome. Ice-formation in the firn zone is of the cold infiltration type. In the pit at the summit of the dome was revealed an alternance of firn and ice-layers, the last firn horizon being encountered at a 16 m depth.

With water infiltration into the firn is related an additional heating of the glacier. Everywhere in the firn zone the temperature of the ice at the upper limit of the constant temperatures' zone is by 10° higher than the air-temperature, and it equals approximately -3° . Outside the zone of firn alimentation, where infiltration is absent, the ice temperature remains cold and approximately equal to air temperature. Cold hard ice frames as a horse-shoe the warm plastic kernel in the central and eastern parts of the dome; this contributes to the ice-flow in eastern direction via the Obrutchev glacier. On the Churlianis dome, where the firn zone is absent, ice temperatures is everywhere almost equal and it varies near -10° and -12° in dependence of flow conditions.

Even relatively small differences in climatic conditions can lead to the forming, in one and the same glacier region, of glaciers that differ greatly as to ice formation type, temperature regime and other aspects of vital activity. And what is more, similar differences arise even inside the limits of one glacier.

Nevertheless, the high sensivity of glaciers against climate variations in space do not coincide with their reaction upon climate variation in time. A definite type of glacier alimentation has considerable inertia; this is due to the mechanism of ice formation.

RÉSUMÉ

Les dômes de glace sont un des plus importants éléments des champs de glaces de la Terre de François-Joseph. Leur majorité fait partie de la région d'alimentation de glace correspond à l'alimentation des glaciers du type Baffin). Comme but du rapport présent sert l'éclaircissement des raisons, portant sur les différences entre les types d'alimentation des dômes et la caractéristique de l'activité vitale des dômes à l'alimentation de névé. En guise d'exemple de ces derniers fut choisi le dôme Jackson sur l'île Hoacker, qui est comparé au dôme Tchourlianis sur la même île, qui s'alimente selon le type de glace.

Selon les données des levés pour measurements de neige, la quantité des dépôts pareille à celle du dôme Tchourlianis et est égale à 300-350 mm dans la couche d'eau. D'autre part, la grande hauteur, en comparaison au dôme Tchourlianis et

l'éloignement de la mer amènent à des conditions moins favorables pour l'ablation. Par la différente valeur d'ablation s'expliquent les différences des types d'alimentation sur les sommets des deux dômes.

Les régimes de température et de radiation des parties périphériques du dôme Jackson contribuent en grande mesure à l'ablation et ceci, à son tour, se reflète sur les types de la formation de la glace. Sur le dôme Jackson, l'on distingue trois zones verticales : celle d'alimentation de névé, celle d'alimentation de glace et celle d'ablation. L'altitude de la limite du névé est 325-390 m, la hauteur de la ligne d'équilibre — 270 m au-dessus du niveau de la mer.

La zone du névé sur le versant oriental du dôme Jackson s'adjoint au bassin de névé du glacier Obrouchev, qui descend plus bas que la limite du névé sur le dôme. La formation de la glace dans la zone de névé est du type froid d'infiltration. Dans le puits au sommet du dôme, l'on a révélé une alternance des couches de névé et de glace, le dernier horizon du névé étant rencontré à une profondeur de 16 m.

A l'infiltration d'eau dans le névé est liée une chauffe complémentaire du glacier. Partout dans la zone du névé, la température de la glace près de la limite supérieure de la zone des températures constantes est par 10° plus haute que la température de l'air et elle égale approximativement -3° . En dehors de la zone d'alimentation de névé, où l'infiltration d'eau est absente, la température de la glace reste froide et coïncide à peu près à la température de l'air. Une glace froide et solide encadre en manière de fer-à-cheval le noyau tiède et plastique dans les parties centrale et orientale du dôme; cela contribue au déversement de la glace en direction orientale à travers le glacier-Obrouchev. Sur le dôme Tchourlianis, où la zone de névé est absente, la température de la glace est partout à peu près pareille et varie entre -10° et -12° en fonction des conditions de déversement.

Même des différences relativement petites entre les conditions climatiques peuvent amener à une formation dans la même région glaciaire de glaciers, qui se distinguent essentiellement selon le type de formation de glace, le régime de température et les autres aspects d'activité vitale. De plus, des différences pareilles surviennent même dans les limites d'un glacier unique.

Néanmoins, la haute sensibilité des glaciers aux variations du climat en espace n'est point en conformité à leur réaction contre la modification du climat en temps. Un type défini d'alimentation des glaciers possède une inertie considérable, et ceci peut être expliqué par le mécanisme de la formation de la glace.

Franz Josef Land is the northernmost archipelago of the Soviet Arctic, a region with developed sheet glacier. The glaciers occupy about 87 per cent of the surface. On most of the islands the ice sheet is made up of several ice caps and short outlet glaciers. The caps are inter-connected by ice isthmuses. In places the glaciers are 300 metres thick. Only rarely do bare rocks and sectors of basalt plateaus emerge from beneath them.

The investigations made by P.A. Shumsky and G.A. Avsyuk (G.A. Avsyuk, 1959) showed that ice caps with both ice (*) and firn nourishment on the summits are very widespread in the archipelago. The vital activity of both types was studied by the Glaciological Expedition of the Institute of Geography of the Academy of Sciences of the U.S.S.R., which worked under the I.G.Y. programme in which the author took part.

In this paper an attempt is made to establish the reasons for the difference in the type of nourishment of the above ice caps and to describe the regime of the firn-nourished caps on the example of one of them—the Jackson cap on Hooker Island.

The first question may be settled by a comparison of the climatic conditions on the summits of the Jackson cap and on the ice-nourished Churlianis cap on the same (Hooker) island.

The types of nourishment of both caps are investigated from cross sections of the firn-ice layers.

Churlianis cap, which has an absolute elevation of 360 metres, is situated in the northern part of the island. Jackson cap, situated to the south-west of Churlianis

(*) Ice nourishment is a term, used by P.A. Shumsky in 1949 for infiltration-congelation formation of ice (P.A. Shumsky, 1955) and corresponds to the nourishment of the Baffin type of glacier (Baird, 1952).

cap and farther away from the sea, is the biggest of the caps on the island and rises to a height of almost 450 metres. The summits of both caps are rather flat, Jackson cap's area being nearly 80 km² and the Churlianis 6 km².

At the moment we have for Hooker Island a long series of meteorological observations taken at the Polar base in Calm Bay close to sea level, two-year observation by an expedition on Churlianis cap and only episodic observations for Jackson cap. For that reason in order to compute the mean temperature on Jackson cap we were compelled to use the vertical temperature gradient computed on the basis of the difference of temperature between Calm Bay and Churlianis cap. We reduced all the values obtained to a series of observations in Calm Bay (Table 1).

TABLE 1

Air temperature at various point on Hooker Island

Location	Elevation, metres	Mean annual temperature:		Mean July temperature:	
		Sept. '57 to Aug. '59	for 1930-59	for 1958 and 1959	for 1930-59
Calm Bay	16	- 12.1	- 10.2	2.0	1.2
Churlianis cap	351	- 14.0	- 12.1	0.0	- 0.8
Jackson cap	445	- 14.5	- 12.6	- 0.5	- 1.3

These data show that on the summit of Jackson cap there is an "eternal frost" (Ef) climate after the Koeppen classification, and even during the high mean July temperature that prevailed in 1958 and 1959 it remained negative. Between the temperatures of the summits of the Jackson and Churlianis caps, which are correspondingly firm and ice nourished, there is a mean difference of 0.5°C. The data obtained directly from episodic observations in April and June 1959 coincide with the results of the computation. During these observations the mean temperature on Jackson cap was found to be 0.4°C lower than on Churlianis cap. The very substantial difference in the sum of summer positive temperatures is especially consequential. With a difference of 0.5°C in the summer temperatures on these caps the sums will vary by approximately 1/3.

The positive sum of the radiation balance must likewise be small on the summit of Jackson cap. Here the magnitude of the net radiation is reduced by fogs more than in other places of the island's ice sheet. In Calm Bay the fogs in July-August 1958 lasted 5.4% of the time, on Churlianis cap 60.3%, and on Jackson cap 90%. It should be mentioned that according to observations fogs on Churlianis cap reduced the magnitude of the net short-wave radiation by 60%. The reduction of the absorbed radiational warmth is also influenced by the relatively high albedo in July. On Jackson cap there is an absence of melting ice and pools of water even in July, while these occur on the summit of Churlianis cap and have an albedo of 0.30-0.50. The wet snow, of which the surface of the Jackson cap consists in July, has an albedo of 0.60-0.70.

Summer fogs not only reduce the radiation balance but, increasing the humidity of the air, also greatly reduce the possibility of evaporation.

The temperature and radiation regimes on the periphery of Jackson cap, especially in the low-lying western part, are essentially different from that of the summit. The air temperature here is much higher, and, because of the absence of fogs the uptake of radiation heat is greater than at the summit by at least a third. The albedo value drops because of pollution of the surface by fine debris.

Snow surveys conducted by us showed that the quantity of solid precipitation is approximately the same on Jackson and Churlianis caps and in the period of accumulation in 1958-59 amounted to 340 mm on the summit of Jackson cap. These values are apparently somewhat higher than the mean annual figure because there was very intensive primary circulation in the winter and spring of 1958/59.

The distribution of precipitation on the surface of the caps is determined by wind-borne snow, whose role varies with the difference in terrain on the cap. The thickness of the snow covering on the bulging parts of the caps depends on the steepness of the slopes. For example, on the western slope on Jackson cap in April 1959 the largest quantity of snow was found on the flat summit (85-100 cm) and at the foot of the slope (100-130 cm), and in the middle steep section of the slopes the layer of snow was thinned down to 50-60 cm. The accumulation of snow of the western edges of the caps is facilitated by the predominant easterly and south-easterly winds. A great deal of snow accumulates in the outflow basins of the outlet glaciers fed by the caps.

The heterogenousness of the climatic conditions and of the accumulation of precipitation in the different parts of the cap is due to the presence on it of several ice formation zones. The whole of the upper part of Jackson cap is in the zone of firn nourishment. There is no such zone on Churlianis cap. The height of the firn line fluctuates on the different sides of Jackson cap. On the flat northern and western slopes, where the wind drift is not big, the firn line is 325-340 metres high, and on the steep southern slope it rises to 390 metres. The eastern slope of Jackson cap is in a special position. There the firn outflow basins of Obruchev glacier abut the sub-summit region of firn nourishment. Old firn has been observed by us in all pits on the eastern slope.

Firn nourishment on the summit of Jackson cap is the result of a different water and heat balance than on Churlianis cap. For example, on the summit of Churlianis cap the mass balance was negative in 1959. At the same time, on Jackson cap this balance was positive and almost equalled (with the exception of expenditure through evaporation) to the annual sum of precipitation, because even during this exceptionally warm year there was no run-off water above an elevation of 380-400 m. All the melt water from the surface was absorbed by the layers of snow and firn beneath the surface. On the summit of Jackson cap on August 14, 1959 there was still 25 cm of snow which had not only not melted but had also not been saturated with water.

Thus, the existence of zones of firn nourishment on the tall caps of Franz Josef Land are due to the drop of temperature with altitude, the reduction in the absorption of solar radiation, the absence of run-off melt water and the weaker evaporation compared with ice-nourished caps. The features of the accumulation of snow influence only the fluctuation of the height of the firn line.

With the exception of the eastern slope, the whole of the Jackson cupola below the firn-nourished zone abuts the zone of ice nourishment. Vertically the latter zone is 70-100 metres long. This zone also embraces the summit of Churlianis cap. The equilibrium line passes through all the slopes of Jackson cap, except the eastern slope, at an altitude of 270 metres. Below that is the ablation zone. Judging from the observations made on Churlianis cap, the value of the negative mass balance on the

caps of Hooker Island may, in the most favourable places, attain a layer that would be the equivalent of 300-500 mm of water. In any case, the glacial caps of Franz Josef Land are not solely regions of nourishment.

The mechanism of the formation of ice on firn-nourished caps in the archipelago is of especial interest. Pits and bores on the summit of Jackson cap (Fig. 1) reveal a typical cross-section of a cold infiltration zone in conformity with the classification by P.A. Shumsky (1955). There is an analogous cross-section, for example, in the

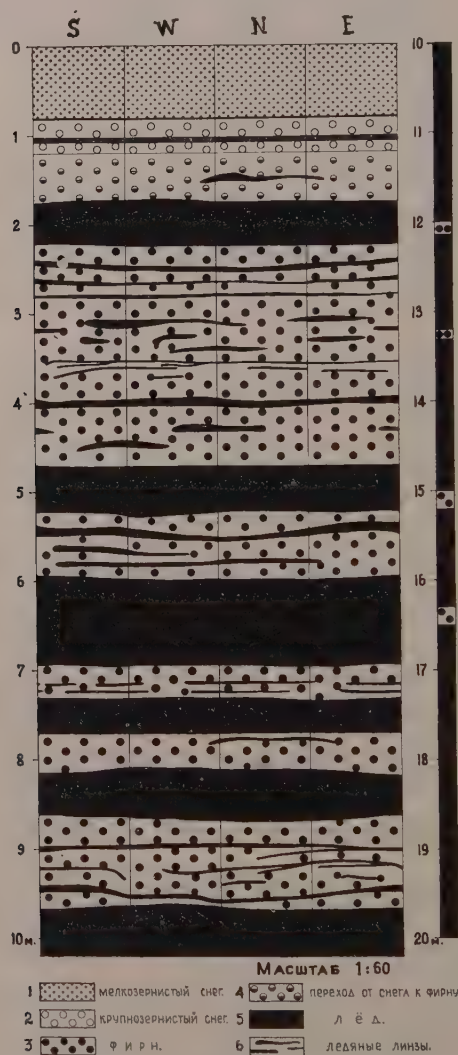


Fig. 1 — Cross section of the firn-ice layer in a pit and borehole on the summit of Jackson cap. The walls of the pit are oriented towards the countries of the world. The borehole was drilled at the bottom of the pit. 1. Fine-granular snow. 2. Large-granular snow. 3. Firn. 4. Transition from snow to firn. 5. Ice. 6. Ice lenses.

snow-firn layer of North-East Land (Glen, 1939) and in the upper regions of Nova Zemlya ice sheet (Chizhov 1961). There layers of ice alternate with layers of firn, and their correlation changes with depth. Firn predominates in the top 10 metres, while between 10 and 20 metres it meets only thin strata, the lowest of which was found at a depth of 16.5 metres. The thickness of the firn levels also drop with depth. The total thickness of the firn is 7.3 metres.

The size of the firn grains is approximately the same throughout the layer. With depth the firn gradually grows more compact, this being shown by the changes of its physical properties with depth (Fig. 2). The firn goes through two stages of development. In the upper 4 m layers freezing of percolating waters and more compact packing of the grains is observed. Below the four-metre line, packing ceases and plastic deformation of the compression of the crystals becomes noticeable which leads to a reduction of the internal porosity and to a decrease of the general process of growth of density. In the closed pores the air pressure, which until then had increased slowly, now grew rapidly, reaching two atmospheres at a depth of eight metres. An interesting identity of data was obtained during observations carried out at "Eismitte" in entirely different conditions of the recrystallization zone (Sorge, 1935).

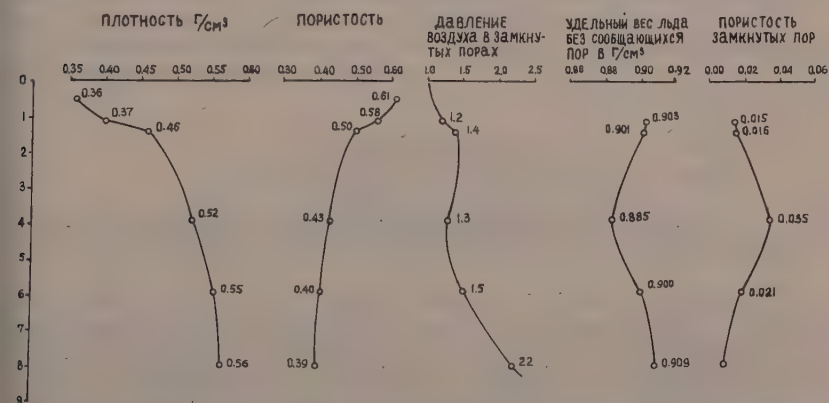


Fig. 2 — Changes in the physical properties of snow and firn with depth on the top of Jackson cap (density, porosity, air-pressure in the bubbles, specific weight without open pores, porosity of closed pores).

The process by which the firn becomes compact does not, however, reach the stage where ice is formed. Ice forms through infiltration when water penetrates into the firn layer. This is testified to by: 1) the character of the ice crystals; 2) the occurrence in it of circular bubbles originating from the air dissolved in water; 3) growth of the volume of the closed pores compared with the layers of firn lying above them. At a depth of eight metres in the firn the specific volume of these pores is 0.011, in the ice at a depth of 9.5 metres it is 0.026; 4) Drop of the air pressure in the closed pores in the ice compared with the layers of firn above them. The fact that melt water penetrates deep into the firn through cracks in the ice strata is confirmed by the faster warming up of the firn strata compared with the ice through discharge of heat by the former when part of the water freezes. This is shown by a comparison of the temperature curves for June 22 and 28, 1959 (Fig. 3) at a five-metre level (in the ice) and in the firn levels above and below it. The absence of other similar bulges in the curve for June 28 is explained simply by the inadequate frequency of observation levels. The numerous strata and lenses of ice are explained by the delay and freezing of the infiltrating water above the ice crusts in the firn.

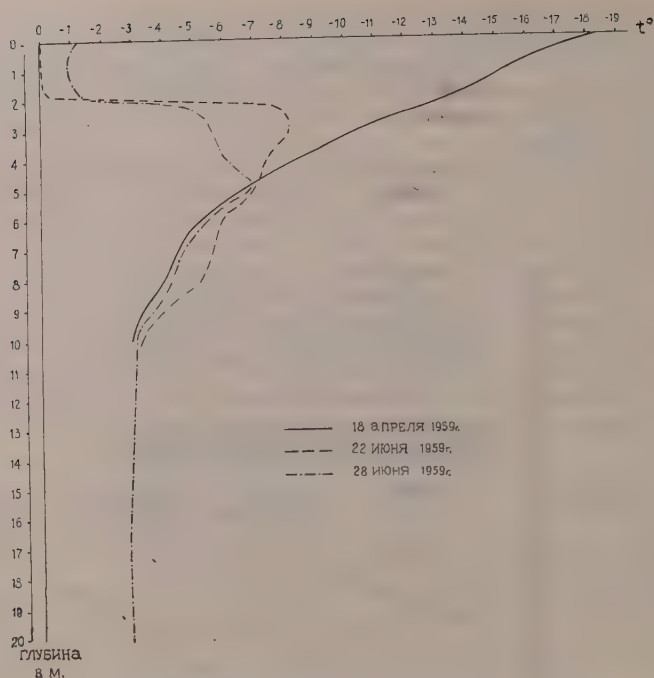


Fig. 3 — Changes in the temperature of the ice and firn with depth on the top of Jackson cap.

The presence of a firn cover in the upper zone of the cap leaves an imprint on the temperature regime through the layer beneath this cover. The heat discharged when the water penetrating into the firn freezes prevents the ice from cooling. A similar effect of the firn cover was noticed by many explorers on different glaciers of the world.

On Churlianis cap where there is no firn the temperature of the ice at a depth of 15 metres was -9.6°C throughout the whole of 1959 (measurements taken by N.G. Razymeiko) and was only 2.5°C above the mean annual temperature of the air, but on the top of Jackson cap the temperature of the ice at the same level was only -3°C in April and in June, thus differing from the mean annual temperature by almost 10°C .

Taking the geothermal stage in the ice as 30-50 metres, we get an ice core with a temperature of 0°C at a depth of 100-150 metres within the Jackson cap.

The 0°C isotherm possibly lies lower than that, while the temperature of the entire thickness of ice is -3°C . Only the very lowest layers of the ice are heated by terrestrial warmth to a temperature of 0°C and partially melt, simultaneously absorbing the heat that in the case there is no melting here would have gone to heat the higher layers through heat-conductivity. We find it highly probable that there is melt ground under the central part of the cap.

Thus, in any case, the central part of the cap is made up of relatively warm ice. The fact that there is a firn cover on the eastern slope of the cap abutting the firn basins of Obruchev glacier must lead to the formation of a tongue of warm ice extending far to the side of this glacier, which is evidently also warm.

Judging from the measurements made on Churlianis cap we expect that where the cap is devoid of a firn cover (zones of ice nourishment and ablation) the tempe-

perature of the ice must be much lower, close to the mean annual temperature of the air. The schematic distribution of warm and cold ice in the body of Jackson cap is shown on the drawing (Fig. 4).

As a matter of fact, firstly, a considerable quantity of low-temperature ice comes into the ice-nourishment zone in the body of the cap. Judging from the measurements taken on Churlianis cap, which show that in this part of the cap the ice moves at a slow rate not exceeding a few metres a year, an annual addition of even 5 cm of ice creates a layer of ice several tens of metres thick in the 1-2-kilometre wide lower part of the zone.

Secondly, when the warm ice emerges from beneath the firn cover it begins to cool through heat-conductivity, because this cooling is now not hindered by the discharge of heat as happens when infiltrating water freezes. At the above-mentioned slow rate of movement, the temperature of the ice drops sharply already a few hundred metres from the edge of the firn cover. The cooling is facilitated by the lesser thickness of the ice near the periphery of the cap. Practically speaking, throughout the entire period of the modern climate, the ice in the zone of ablation and on the periphery of the ice zone remained outside the heating action of the firn. Its temperature regime is determined by the processes of heat exchange with the air and the underlying rocks, and does not originate from the warm ice of the firn zone.

Consequently, in the central part of the isolated firn-nourished cap there is warm plastic ice held in a ring of hard ice (Fig. 4).

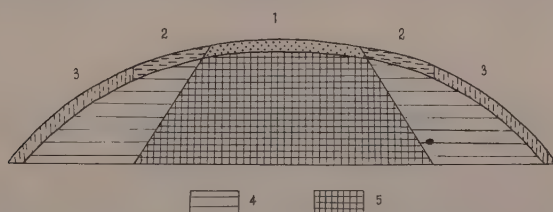


Fig. 4 — Scheme of the distribution of «Warm» and «cold» ice in the body of a firn-nourished glacial cap (cross-section). 1. Zone of firn nourishment. 2. Zone of ice nourishment. 3. Zone of ablation. 4. «Warm» ice. 5. «Cold» ice.

On Jackson cap the picture is made more complicated by the presence of the above-mentioned east-oriented warm tongue that has taken shape through the descent of ice through the outlet Obruchev glacier, in whose firn basin the ice is likewise safeguarded against cooling. Meanwhile the cold hard ice of the cap forms not a ring but a horseshoe with its open side facing east.

No observations have been made so far of the movement of ice on Jackson cap, but indirect data allow us to arrive at some surmises with regard to this movement. The first is based on the distribution of temperature in the body of the cupola. The distribution of hard and plastic masses of ice makes for the eastern orientation of the main current of ice. In this case, the relatively thin (about 10 metres) layer of firn must create a powerful run-off of «warm» plastic ice among the hard ice «shores» Indeed, judging by the piling of sea ice in front of the glaciers (Viset, 1928), Obruchev glacier is the most active on the Hooker Island.

There is absolutely no westward run-off of ice. The entire western slope is devoid of cracks and for a whole series of reasons is composed of dead ice. In the north and south Jackson cap nourishes the fairly active Yuri, Avsyuk and Kirov outlet glaciers.

The expenditure of ice through movement and melting in the ablation zone exceeds the formation of new ice on the top of the cap. This is testified to by the repeated

observations made by M.G. Grosswald of the edge of the cap and of the position of modern marginal canals. Judging by the distance between the latter, the Jackson cap retreats by an average of 5 metres annually.

From what has been said above it follows that even relatively small differences in climatic conditions and in the conditions of accumulation can lead in one and the same glacial area to essentially different types of ice formation and temperature regimes in the ice, and, consequently, to other differences in the activity of the ice. Moreover such differences may arise even in one glacier.

Within definite climatic limits, the temperature of the ice may drop when the ice passes to a zone with a higher air temperature, and conversely. This paradox is explained by the role played by the hidden heat of melting, and is observed, it goes without saying, only when melting reaches a sufficient intensity during the period of ablation.

Lastly, it should be borne in mind that the high sensitivity of glaciers to climatic changes in space does not correspond to their reaction to climatic changes in time. A definite type of nourishment of glaciers with all the consequences arising from it possesses a considerable inertia, which is explained by the mechanism of the formation of ice.

In conclusion I should like to express my gratitude to my comrades in the expedition— V.A. Markin, L.D. Bazanov, N.G. Razumeiko, M.G. Grosswald—who together with me carried out field observations on Jackson cap in conditions made difficult by isolation from the expedition's base.

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STUDIES OF FOUR GLACIERS IN GREENLAND

B. FRISTRUP (Danemark)

SUMMARY

Glaciological investigations were carried out in Greenland in 1956-58 as part of Danish contribution to IGY under direction of the present author.

The main programme of the investigations was to study the glacier types in relation to geomorphology and climatology of Greenland. Four special selected glaciers were investigated as representative for particular geographical provinces, all being local glaciers outside the Greenland Ice Cap, and all of medium or small size; it was presumed that small glaciers will react more sensitively to less pronounced changes of climate.

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The location and size of the glaciers is given in table 1 and in fig. 1.

TABLE 1

Locality	Area	Glacier type	Climate
Hurlbut Gletscher 77°23'30'' lat. N. 67°57' long. W.	188,0 km ²	Glacier Cap	high arctic dry
Sermikavsak 71°11' lat. N. 53°03' long. W.	21,6 —	Valley Glacier	arctic humide
Napassorssuaq Gletscher 60°18' lat. N. 45°13' long. W.	2,1 —	Valley Glacier Greenland Type	subarctic humide
Mitdluagkat Gletscher 65°41' lat. N. 37°54' long. W.	36,4 —	Transection Gl.	arctic heavy snow fall

According to the classification proposed by Hans Ahlmann (1948) a calculation of the different height intervals was made, the curve is shown on fig. 2. The northernmost of the selected glaciers: Hurlbut Gletscher is a glacier cap, and that type of glacier is dominant for the whole North Greenland with exception of the outlet glaciers from the Inlandice. Valley glaciers occurs almost only on the Cape York Peninsula and in the most northwestern part of Pearyland belonging to the Caledonian folding ranges.



Fig. 1 — Locations of the glaciological stations. 1) Hurlbut Gletscher, 2) Sermikavsak, 3) Napassorssuaq Gletscher, 4) Mitdluagkat Gletscher.

The valley glacier is dominant in most part of West Greenland but also other glacier types can be found especially characteristic being the cwm's in the basalt region round Disko Bay, especially on Disko Island. The southernmost of the four glaciers «Napassorssuaq Gletscher is also a valley glacier, but as seen by fig. 2 the curve of the different height intervals in reality is not a typical valley glacier curve, but represents a transition between Ahlmann's valley glacier IV and the piedmont glacier. According to the general morphology of the glacier it is quite evident, that this glacier is a typical valley glacier for Greenland, and a similiar glacier will be found in many places in West Greenland; the name Greenland valley glacier type as the fifth of the Ahlmann glacier types is therefore proposed. The explanation of the difference from the normal valley type is that the lower marginal part of the glacier at the present time is dead ice and is merely a relict of the former greater extension of the arctive glacier but still morphologically a part of the glacier. According to the investigations in North Greenland as pointed out below in this paper, the piedmont glacier or the tongue formed glacier with expanded lower part has previously been typical for that part of Greenland and possibly also for many regions in South Greenland too. The fourth glacier «Mitdluagkat Gletscher» is a transection glacier which is a typical glacier type for the strong glacierized South East Greenland region with heavy snowfall.

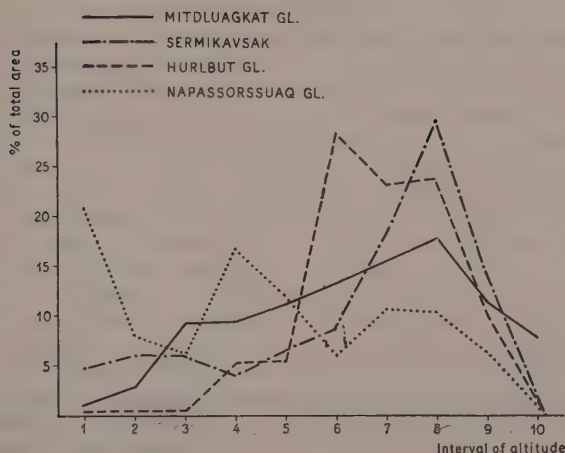


Fig. 2 — Curves of frequencies of different heights intervals for the four investigated glaciers.

Glacial-meteorological investigations were conducted by all the stations, and it was found, that radiation weather was dominant at all the stations. The results of the station «Sermikavsak» have been published by Hans Kuhlman (1959), the results from the other stations are under preparation. According to Kuhlman radiation weather was found in 61% of all observations, overcast was found in 17,7% and foehn weather only in 6,9%. Special investigations concerning temperature and wind profiles above the ice have been conducted, and gravity wind (katabatic wind) was found in 78% of the observations, according to the results round 80% of the ablation was found to be due to radiation. While the foehn conditions are of great importance for the snow melting in spring, the investigations here carried out indicate that the importance of the foehn for the summer melting on the glaciers has been over-emphasized for most of the Greenland Glaciers.

The ablation were measured along bamboo poles drilled 2 to 3 meters down in the ice. At some of the poles there were problems with meltwater in the drill hole during the last periode of the ablation season. The methode of measuring the ablation along poles is a very rough one and gives only a satisfactorily result if measurements are taken along a great number of poles, and even then the results can be rather confusing, as the melting is very strongly influenced by the surface of the ice and by the content of impurities in the ice. For the North Greenland glaciers where the amount of dust and pebbles is rather small it is not of great importance, but especially for the West Greenland glaciers the problem is rather serious. A special test of ablation from different surfaces with different amount of stones, pebbles and imporities were therefore conducted by Kuhlman on Sermikavsak as seen from the table 2 the ablation in the same region with the same exposure to the sun and to the wind and at the same altitude above sea level varied between 0.6 and 1.6 relative ablation, the ablation from pure ice beeing 1.0.

It will be advantageous if the readings from different glaciers could be standardized. Only readings from ice surfaces of a same character can be compared, therefore it will in many cases be necessary to use an artificial surface made by digging and scraping and even ice surface out from the uneven glacier surface.

TABLE 2

gm. stones/cm ²	Relative ablation	
	18-19/7	19-20/7
0	1.0	1.0
1	1.2	1.3
1.9	1.35	1.6
4.9	0.85	0.8
8.6	0.7	0.6

The ablation during the summer time is a rather equal function of time as seen by fig. 3. According to the rather little difference in average summer temperature the daily ablation does not varie much for most of the stations. The length of the ablation season varies rather much with more than two months from south to north.

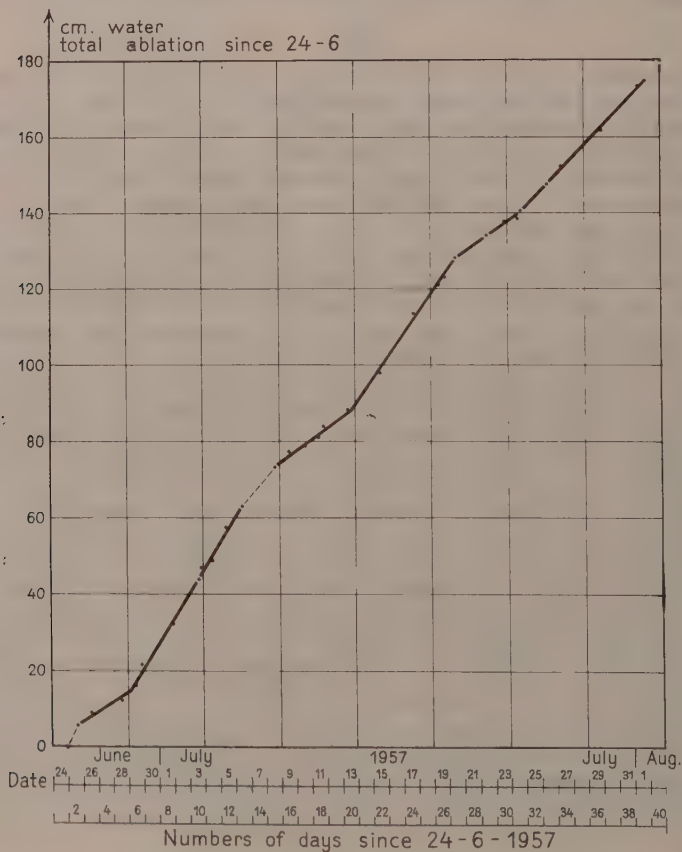


Fig. 3 — Ablation in July at Sermikavsak according to Kuhlman.

Runoff studies were conducted by the Mitdluagkat Gletscher on Angmagssalik Island, and a preliminary report has been published by Hans Valeur Larsen (1959); a typical diurnal variation of the discharge was found with maximums at 17-18 hrs lokal time, the discharge varies between 2 and 4 cbm/sec. Of special importance for the runoff from the Mitdluagkat Gletscher is the occurrence of tapings of ice dammed lakes. By such catastrophic outflow, the discharge was of the same form as found by the great «jökullhlaup» on Iceland, that means a rather gentle increase to maximum and then a sudden drop to nearly normal level. The water from the drained off lakes appeared not in the glacier port at the front but far up on the lateral drainage channels and from there following the glacier margin down to the river. In connection with the drainage of the ice dammed lakes big subglacial tunnels were found; most of the water was drained through tunnels in the ice and not in the contactzone between ice and bedrock, the collapse of such tunnels gave rise to big calderons found several places in the ice, the tapping took especially place in late summer just at the end of the ablation season, and was most probably caused by hydrostatic lifting of the ice damming the lake in connection with heavy rainfall and high water level in the lakes. The process is very typical and these catastrophic situations are a normal event in the life of the glacier and most probably the collapse of the ice tunnels with their amount of stones, clay and pebbles may give rise to some of the occurrences of dirt bands and the amount of stones at the glacier front, which are difficult to explain otherways.

Occurrence of superimposed ice was found to be very important, especially at Hurlbut Gletscher, as in some years the firn area may be vanishing, and melting took place all over the glacier, the glacier so was nourished in the same way as described by P.D. Baird (1952) for Barnes Ice Cap on Baffin Island, E.G. by the refreezing of melt-water and water-percolated snow and ice. The occurrence of superimposed ice has a very wide distribution on the Greenland glaciers, meaning that the zone between the firnline and the equilibrium line is very broad.

According to the present climatic conditions in Greenland all glaciers were found to be retreating at the present moment but the retreat of the fronts was of a quite different magnitude, not only because of the area of the glacier but also in relation to the different geographical positions.

The southernmost of the investigated glaciers was for the first time visited in 1894 and since then the glacier front had retreated 200 meters up to 1951 and the retreat has continued. In 1957 the total retreat was 350 meters which gives us a rather fast retreat during the last years. The Sermikavsak has since 1934-1953 retreated round 600-700 meters, and from 1953 to 1957 the retreat was 150 meters giving an annual withdrawal of 34-38 meters. Hurlbut Gletscher has a glacier tongue descending to the sea at Inglefield Bredning (Fjord) and photographs from 1939 compared with the present photos shows us a rather insignificant retreat most probably less than 5 meters pr. year but the glacier front and the lower part is more narrow than before the lateral part at some places being dead ice that being in good accordance with the results obtained by studying the old moraine systems which in spite of some destructions by solifluctions indicate that the glacier previously has been of piedmont form with the ice from a rather narrow valley parts spreading radially to the sides. This is typical for many of the glaciers in North Greenland and good examples can still be found in many places, however, they most probably are relicts from a time with another climate for the glaciation. A very great retreat has been found on the Mitdluagkat Gletscher. At a visit in 1933 K. Milthers took some phototheodolite exposures and the photos compared with our exposures taken from the same points show a retreat of considerable dimension. The total withdrawal during 25 years is about 400-500 meters, and formations of new nunataks which now divide the glacier tongue into two lobes emerging below the nunataks is very impressing, still new nunataks are under development. The annual withdrawal of the glacier front has been round

16-20 meters. A retreat of a same order of magnitude was found on several other glaciers having been visited in the district.

Temperature measurements down to 15 meters below the surface were taken at regular intervals during the summer, some few readings were taken also in the winter on Hurlbut Gletscher. Some of the results will be evident from fig. 4 and fig. 5. It will be seen that the yearly temperature variation continued down in the depth of

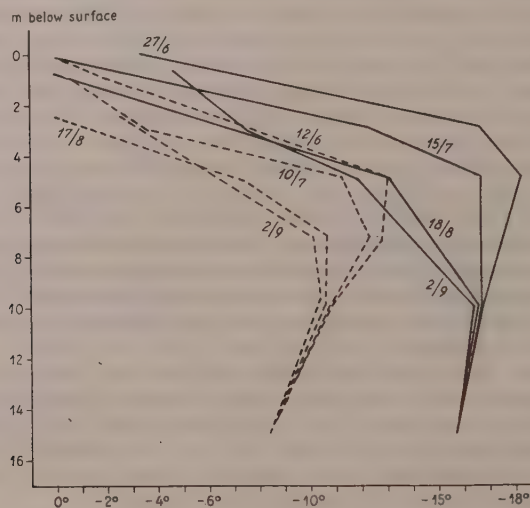


Fig. 4 — The temperature variation during the summer time at two different stations on the Hurlbut Gletscher. The curve to the right is from the station on the Ice Cap near the highest point of the glacier, and the curve to the left is from the glacier tongue descending to Inglefield Bredning.

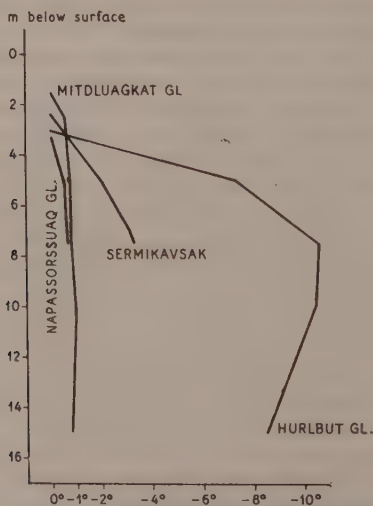


Fig. 5 — The ice temperature at the four investigated glaciers in August.

10-12 meters, in which the temperature is nearly constant, and even in September the cold from the previous winter is still present in the deeper part of the ice. There is a good relation between the ice temperature and the altitude above sea level as demonstrated by the two stations from Hurlbut Gletscher in fig. 5. From fig. 4 it will be seen that there is a great geographical variation of the temperature in the ice of the different glaciers. All the stations being nearly at the same altitude above sea level and the two glaciers Napassorssuaq and Mitdluagkat Gletscher have a temperature in August which is very near the melting point, the temperature may only vary between zero and -1° down to the depth of 15 meters. The Hurlbut Gletscher is a cold gletscher with temperature at -16° in the depth of 15 meters in the highest part and at -8° in the lower part of the glacier tongue while the Sermikavsak seems to be a transition form. This is in good accordance with the difference of the geographical latitude. A similar relation between temperature in the ice and geographical latitude has been demonstrated by Carl S. Benson (1959) for the Greenland Ice Cap. According to Ahlmanns definition of glaciers related to temperatures the Napassorssuaq and Mitdluagkat Gletscher may be considered as typical temperated glaciers belonging to the same type as the Scandinavian and the Iceland glaciers, while the North Greenland glaciers are polar glaciers, as a certain amount of melting takes place. Studies from Pearyland (Fristrup 1951) shows that the ice caps in that region also have a summer melting. Therefore the only part of the Greenland Ice Cap and probably one or two of the largest glacier caps in North Greenland may be considered as real high polar glaciers while all the rest may be subpolar if not temperated glaciers.

As the yearly temperature variations only affect the upper 10-12 meters of the ice it will be seen that the cold glaciers in North Greenland will react very slowly to changes of climate as far as the temperature concerns but will be very sensitive to changes in precipitation. As shown by Diamond (1956, 1958) there has been a slight decrease in annual amount of precipitation during the years 1920-1954 while we at the same time have found retreat of the glaciers. The cold glacier temperatures of the North Greenland glaciers may possibly be the explanation of the question why the northern glaciers according to Lauge Koch (1928), Fristrup (1952) and others seems to have started the withdrawal of the fronts much later than the glaciers in South and Central Greenland, and as seen from the above mentioned investigations the annual withdrawal at the present moment is less for the North Greenland glaciers than for the glaciers in South Greenland.

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LES GLACIERS ENTERRES ET LEUR ROLE MORPHOLOGIQUE

Louis LLIBOUTRY (France)

RÉSUMÉ

L'auteur décrit la névéification rapide de la neige subsistant l'été dans les Andes de Santiago vers 4.000 m. Au pied de certains couloirs ce névé, recouvert chaque année par des éboulis, donne naissance à de petits glaciers enterrés. Les strates de terre de ces glaciers se relèvent très fortement à l'aval. L'auteur suppose que progressivement de tels glaciers enterrés s'enfoncent profondément dans la pente meuble. Ils s'écoulent hors de leur niche en langues qui se transforment ultérieurement en glaciers rocheux.

SUMMARY

The rapid firnification of the snow which remains during the summer at 4.000 m in the Andes of Santiago is described. At the foot of some slopes this firn is covered each summer by rubble and becomes transformed into small buried glaciers. The earth strata of these glaciers emerge at great angles near the front. The author assumes that such buried glaciers progressively sink into the loose material of the slope. They flow from their niche as tongues which in the course of time become rock glaciers.

1. INTRODUCTION

Si l'on convient d'appeler *glacier* toute masse de glace permanente (du moins à l'échelle humaine) d'origine nivale, on peut se demander quelle est la forme initiale, le glacier le plus simple. Il est classiquement admis que c'est un petit névé, soit reste d'une corniche que l'été n'est pas parvenu à fondre, soit tache de vieille neige abritée dans une « niche de nivation ».

Dans la bordure ouest du massif du Nevado Juncal, au N.E. de Santiago du Chili, région longuement étudiée de 1952 à 1956, j'ai observé ces formes initiales et même assisté au développement d'un petit glacier. Dans la partie nord de la concession de la Mine La Disputada de Las Condes, 400 m au Nord-Ouest de la petite Laguna de La Copa, vers 3 720 m une corniche orientée vers le Sud-Ouest s'est formée pendant l'hiver 1953, exceptionnellement neigeux, et n'est pas parvenu à fondre pendant l'été. En avril 1956 cette corniche était devenue un « névé permanent » de 300 m de large et de 100 m dans le sens de la pente.

Or cette région, entre 4 200 m, est pleine de glaciers rocheux, à différents stades de leur évolution. Il y en a 7 ou 8 de bien caractérisés dans un rayon de 3 km autour de la Mine (*). Suivant l'opinion générale, prouvée dans certains cas par l'observation les glaciers rocheux proviennent de petits glaciers de cirque dont la langue était extrêmement riche en débris morainiques. Aussi semble-t-il évident que les cirques occupés par des glaciers rocheux soient les lieux où, à une altitude donnée, les névés persistent de préférence. Or aucun des 4 névés subsistant l'été 1953-54 aux environs de la Mine La Disputada n'occupait l'un de ces cirques, ne se trouvait à la tête d'un glacier rocheux. Il est donc douteux qu'une plaque permanente de névé évolue vers un glacier recouvert, puis un glacier rocheux.

Mais dans ces régions il existe une autre forme initiale de glacier, forme occulte et discernable seulement par un observateur averti : de petits glaciers entièrement recouverts d'éboulis l'été, s'alimentant chaque hiver d'une couche de neige, rapidement transformée en glace. J'appellerai de telles masses de glace d'origine nivale des *glaciers enterrés*. (Je les avais appelées glaciers d'éboulis dans une note à l'Académie des

Sciences ⁽²⁾, mais ce terme a déjà été employé à juste titre par Corte, le cryopédologue argentin, comme synonyme de glacier rocheux.) Ce sont ces glaciers enterrés qui semblent conduire le plus souvent à la formation de petits glaciers rocheux.

On trouvera ci-joint (fig. 1) une photo, prise en février, de trois glaciers enterrés au pied du Cerro Negro. L'on observe également de nombreux glaciers enterrés dans le Cajon Barriga, 9 km au Nord de ceux du Cerro Negro, et sur le versant nord du Nevado Juncal. Plus au Sud, il en existe au Paso Molina (col frontière au droit de Rancagua). C'est évidemment une forme plus difficile à reconnaître qu'un glacier ordinaire ou un glacier rocheux, aussi est-il actuellement impossible de préciser son extension.



Fig. 1 — Versant ouest du Cerro Negro (Andes de Santiago)

Les glaciers enterrés du Cerro Negro se situent entre 4000 et 4100 m, alors que la ligne d'équilibre des glaciers se situe dans la région à 4300 m environ. Ils rappellent donc par leur position basse, au pied d'un couloir qui leur fournit une alimentation en neige abondante, les glaciers de cône d'avalanche étudiés par Dino di Colbertaldo ⁽³⁾. Ils rappellent encore plus les *éboulis avec neige interstratifiée* signalés par Demangeot ⁽⁴⁾ sur les pentes du Brec de Chambeyron, une région des Alpes françaises également à étés secs, et célèbre par ses glaciers rocheux et sols polygonaux. La principale différence est que la transformation de neige en glace est beaucoup plus rapide dans les Andes de Santiago.

2. TRANSFORMATIONS DE LA NEIGE VERS 4000 m, DANS LES ANDES DE SANTIAGO

J'ai décrit par ailleurs le climat des Andes de Santiago ⁽⁵⁾. Vers 4000 m il tombe en hiver et au printemps l'équivalent d'un mètre d'eau ou davantage, sous forme de neige poudreuse, en grosses chutes espacées. De décembre à avril inclus par contre, les précipitations sont négligeables, des semaines ou des mois de beau temps se succèdent sans un seul nuage même en fin d'après-midi. L'air se dessèche rapidement en passant sur des sols échauffés par le soleil et sans végétation. La température de l'air en été, à 4000 m fluctue quotidiennement entre 10 à 15°C vers 14 h et - 3°C à + 2°C à l'aube. Par suite du fort rayonnement nocturne le gel reste intense toutes les nuits.

La gélivation intense et la nature des roches (andésites à grain assez fin, microgranites et granites de contact) expliquent la formation intense d'éboulis, et leur

altération jusqu'au stade de limons. Au dessus des caillasses, s'élèvent vers les crêtes de nombreux couloirs de terre dure dégarnis de neige dès le printemps, d'une pente inadmissible sous nos climats à averses, couloirs qu'on ne peut gravir aisément que crampons aux pieds.

Avec de telles oscillations thermiques la neige de l'hiver se transforme rapidement. Dès l'été elle est devenue du névé, si l'on appelle ainsi une neige qui, malgré sa fusion superficielle aux heures chaudes, peut encore supporter le poids d'un homme à pied. Pour être plus précis, je décrirai deux masses de neige de l'hiver.

1) L'une, qui subsistait en janvier 1955 à 4070 m, à côté des glaciers enterrés du Cerro Negro, sur un sol de terre, de 6×4 m en surface et 1,30 m de haut, avait une densité de $0,60 \pm 0,04$. Le diamètre des grains était 1,5 mm à la base, 1 mm à la surface. Le dessus de la surface était creusé d'alvéoles en nids d'abeilles. La face tournée vers l'Ouest était subverticale et creusée de cannelures verticales. Les arêtes entre les cannelures étaient recouvertes de poussières. Cette concentration des poussières sur les arêtes est due à ce qu'au fur et à mesure de l'ablation les poussières restent en surface et suivent des trajectoires perpendiculaires à la surface. Quant au mécanisme de formation de ces alvéoles et de ces cannelures, déjà décrites par Workman sur les neiges du Karakorum ⁽⁶⁾, il est assez analogue à celui de la formation des pénitents : les radiations du soleil et du sol se concentrent dans les creux et les approfondissent de plus en plus.

Cette masse de neige était formée de couches très homogènes, séparées par des strates de glace correspondant aux périodes de beau temps. La fig. 2 représente une coupe de cette masse de neige suivant la ligne de plus grande pente. On remarquera que les strates se relèvent vers l'aval. Pour l'expliquer on peut invoquer :

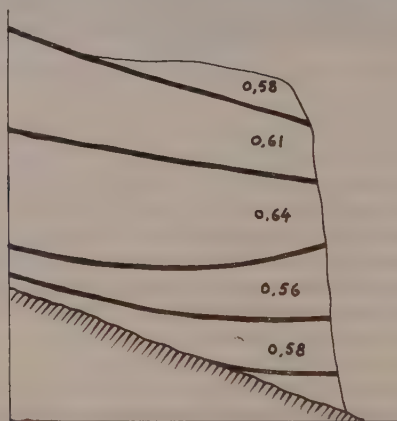


Fig. 2 — Coupe d'une masse de neige observée en janvier 1955, à 4070 m, au pied du Cerro Negro (Andes de Santiago). Les nombres représentent les densités.

- a) Une déposition non uniforme.
- b) Un relèvement des strates dû à une reptation de la neige, avec compression longitudinale.
- c) Une accumulation de l'eau de fonte dans la partie aval, eau qui regèle la nuit, contribuant ainsi à grossir la partie aval au détriment de la partie amont. L'existence d'une première et d'une dernière couches incomplètes; ne subsistant plus qu'à l'aval, me fait préférer cette dernière explication.

2) A 100 m de là, une tache de vieille neige reposait sur de la glace. La névéification était encore plus poussée (densité : 0,62; grains de 2 mm de diamètre) Quant à la glace : a) elle était jeune, ses grains n'ayant que 3 ou 4 mm de diamètre; b) elle provenait directement du névé, comme le prouvaient les nombreuses bulles d'air; c) elle avait été pénétrée par le soleil, chaque bulle étant entourée d'une poche d'eau. Le névé présentait ici des fentes transversales Est-Ouest, dues à une rupture de pente, et donc témoignant d'une reptation de l'ensemble. Ces fentes étaient transformées en trous ovales, «trous méridiens» d'Agassiz, de 1 à 2 m de profondeur.

Dans les Andes de Santiago la neige de l'hiver, rapidement métamorphosée et consolidée, se transforme dès le printemps en pénitents. Le fait est général au dessus de 3500 m au printemps. Mais à mesure que la saison avance, les pénitents fondent et se détruisent dans les régions inférieures, où il n'y a plus de gel nocturne. En plein été on ne trouve des pénitents qu'au dessus de 4300 m, et il n'y en a plus à l'altitude d'environ 4000 m où sont faites ces observations.

Il reste toutefois sur certaines pentes de glace à peu près Nord-Sud une trace des pénitents qui les ont recouvertes, et qui étaient alignés à peu près Est-Ouest, c'est-à-dire transversalement à la pente : la pente est formée de banquettes, semblables aux gradins d'un escalier. La fig. 3, coupe transversale relevée vers 4300 m, en février, sur le côté opposé de la Vallée du Rio Blanco qui nous occupe, montre immédiatement la genèse du phénomène.

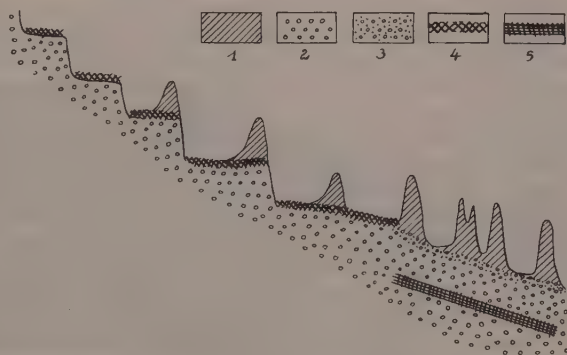


Fig. 3 — Formation de banquettes de glace à partir de pénitents au pied des Puntas del Infernillo (Andes de Santiago). Février 1955, 4300 m, pente nord-est.

- 1 : névé
- 2 : glace à bulles
- 3 : glace limoneuse jaunâtre
- 4 : sol hétérométrique gelé
- 5 : sol gelé avec strates de glace

C'est sur ces masses de neige alvéolée, creusée de trous méridiens, sur ces banquettes et sur des pénitents en voie de disparition que roulent et se déposent des éboulis pendant tout l'été. Aussi la couverture d'éboulis est loin d'être uniforme, ce qui explique les aspects qui vont être décrits ci-dessous.

3. COUPE D'UN GLACIER ENTERRÉ

Les quatre glaciers enterrés, juste au Sud du Cerro Negro, au pied de la falaise, naissent en pleine pente, sans cirque d'alimentation. Ils émergent d'une pente uniforme

d'éboulis de 350 m environ, pour former une langue de 80×150 m environ. Leur pente superficielle diminue de 30° jusqu'à 15° au front, et ils se terminent brusquement vers 3980-4000 m par un talus raide (40 à 45°).

Le tout est uniformément recouvert d'éboulis hétérométriques, où toutes les tailles sont représentées depuis les cailloux de quelques centimètres ou même décimètres jusqu'au limon que peut emporter un courant d'eau non turbulent. 10 à 25% du matériau traverse un tamis de 200 mailles au pouce. Ces éboulis ne rappellent en rien les «caillasses» des Alpes formées exclusivement de cailloux délavés par les averses.

J'ai eu la bonne fortune de pouvoir observer une coupe naturelle d'un glacier enterré, sur les parois d'un petit ravin creusé par l'eau de fonte venant des pentes supérieures. Car creuser le glacier au pic est impossible, à cause de l'altitude qui essouffle rapidement et du nombre de pierres enchâssées dans la glace. La profondeur de la coupe était de 3 m environ (fig. 4).

Dans la partie haute les éboulis superficiels sont épais et stratifiés, des strates épaisses de terre alternant avec 3 à 8 lits minces de gravier. Comme je l'ai signalé ailleurs (7), après avoir assisté à leur genèse, ces dépôts de pente stratifiés résultent de l'alternance d'années normales, pendant lesquelles se déposent des lits de cailloux provenant de la gélivation (le matériel fin, s'il y en a, étant emporté par l'eau de fonte de la neige), et d'années anormalement humides, pendant lesquelles des coulées boueuses en nappe viennent recouvrir ces lits de pierres.

Dans la masse du glacier, la coupe montre que les strates de débris ne sont pas régulières. Elles s'interrompent, se dédoublent, reflétant la surface tourmentée des dernières taches de neige. Il se trouve aussi des poches de débris de forme irrégulière, probablement anciens trous méridiens comblés de débris. *Nulle part l'on n'observe de signe de fort cisaillement, de laminage.*

La glace a des grains de 4 mm de diamètre moyen dans la coupe amont et dans la couche supérieure de la coupe aval. Dans le bas, les grains de glace ont environ 15 mm de diamètre, mais il se trouve aussi des zones de petits grains, de quelques millimètres de diamètre seulement. L'aspect de cette glace ne rappelle en rien celui des lentilles de glace de ségrégation qu'on rencontre dans le permafrost (que l'auteur a eu l'occasion d'observer au Groenland). Par endroit on note, au voisinage d'une strate de débris à peu près horizontale, de la glace à bulles d'air intracrystallines allongées (1 mm de diamètre, 5 à 8 mm de long), comme on en trouve souvent dans la glace froide, au voisinage de la surface libre. Ces bulles ne sont pas allongées à la suite d'un cisaillement de la glace, car les débris voisins n'en portent aucune marque. Je les considère (sans en avoir de preuve, et ce n'est là qu'une suggestion) comme d'anciennes bulles intercrystallines allongées formées dans une glace de regel colonnaire, bulles qui sont devenues intracrystallines à la suite d'une recristallisation de la glace.

Il existait également, au dessus d'une strate de débris, une cavité très aplatie (60 cm de profondeur, 3 cm de haut), dont le sol, boueux, était horizontal; la voûte, verglacée et munie de petits stalactites. Il semblerait qu'il y ait eu tassement des débris dans la poche qu'ils occupaient, et donc qu'ils aient été emprisonnés avec de l'air. En tout cas une telle cavité semble prouver que cette glace n'a jamais été profondément enfouie.

Mais le fait le plus frappant, c'est le pendage de plus en plus fort des strates lorsqu'on descend le glacier. Presque horizontales vers le haut du glacier enterré (et donc émergeant suivant un angle de 15 à 20° avec la surface), elles émergent 100 m plus bas suivant un angle de 45° à 90° . En cette région, près du front, les strates d'un bloc de glace vieille ne font plus qu'un angle de 20° avec la verticale, et sont donc perpendiculaires à la surface.

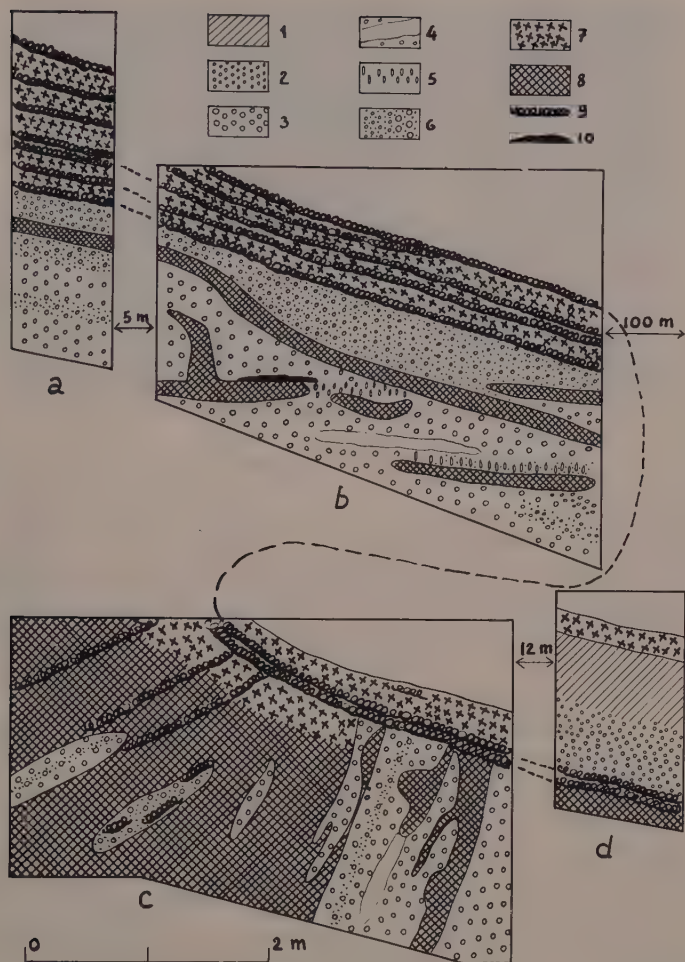


Fig. 4 — Coupe longitudinale naturelle d'un glacier enterré du Cerro Negro

- 1 : névé
- 2 : glace jeune
- 3 : glace vieille bulleuse
- 4 : glace sans bulles
- 5 : glace à bulles allongées verticalement
- 6 : glace limoneuse jaunâtre
- 7 : sol hétérométrique sec
- 8 : sol hétérométrique gelé
- 9 : lit de cailloux
- 10 : cavité

4. ROTATIONAL SLIP, CONTRACTION DU FRONT, AFFOUILLEMENTS

On songe, en voyant ce relèvement des strates au glissement circulaire (rotational slip) cher à Lewis^(*), inspiré des glissements de terrain observés en Mécanique des Sols. A ce propos il convient de faire les remarques suivantes.

1) Un glissement de terrain est un phénomène brusque, dans lequel les forces d'inertie ne sont pas négligeables, et l'équilibre quasi-statique n'est pas réalisé. Aussi l'écoulement peut être différent de celui d'un glacier, qui est un phénomène très lent. Dans un glacier, ces conditions d'équilibre quasi-statique font que le mouvement ne peut pas être plus rapide en profondeur qu'en surface. Tout au plus peut-il être très voisin. C'est le cas pour le glacier de cirque étudié par Mac Call, où Lewis voit à tort un exemple de glissement circulaire.

2) Mais, l'étude de Mac Call le prouve, on peut avoir un très fort relèvement des lignes de courant au front. Certes les solutions données par Nye aux équations de l'écoulement d'un glacier conduisent à des angles d'émergence des lignes de courant assez faibles (inférieurs à 27° dans l'hypothèse du plastique parfait). Seulement Nye suppose que les contraintes sont indépendantes de x (axe dirigé dans la direction de la ligne de pente), c'est-à-dire que le poids du glacier est équilibré par le frottement sur le lit. Pour une petite masse de glace ce peut ne pas être le cas : le frottement sur le lit peut être négligeable. C'est alors un obstacle à l'aval qui retient le glacier.

3) Si Lewis insiste sur ce type d'écoulement, c'est qu'il y voit la cause du creusement des cirques. C'est au contraire parce que le cirque est creusé et que son lit est poli et lisse, que le frottement est devenu négligeable et que le glissement presque circulaire peut avoir lieu.

En résumé un fort relèvement des lignes de courant suppose un lit suffisamment glissant et à l'aval une butée : moraine ou verrou rocheux.

Quel va être le comportement d'une strate superficielle au cours d'un tel déplacement ? L'angle très aigu qu'elle forme avec les lignes de courant ne va pas changer dans la partie haute. Puis vers le front, là où les lignes de courant divergent fortement et émergent à la surface, cet angle va encore diminuer. Simultanément la strate sera étirée et laminée.

Dans le cas du glacier rocheux examiné :

1) Au lieu d'un lit rocheux glissant, il s'agit vraisemblablement d'éboulis meubles. Le frottement du glacier doit également être très faible.

2) A l'aval la glace disparaît et il ne subsiste qu'une langue de moraine, formant une butée.

3) Mais les strates observées ne présentent aucun indice d'étirement et de laminage. La glace chargée de débris semble avoir pivoté en bloc, sans se déformer. Le mouvement a donc été différent de celui observé par Mac Call.

4) Mais il ne s'agit pas non plus d'un glissement de terrain brusque, car on n'aperçoit à l'amont aucune fente transversale, aucun décrochement, aucune banquette.

Nous allons émettre une hypothèse qui rendrait compte de l'aspect observé : c'est que les éboulis formant le lit du glacier se sont progressivement creusés sous lui par suite de l'élimination des parties fines par l'eau de fonte. Il y aurait eu affouillement du lit meuble, faisant basculer la glace près du front vers l'arrière. Le glacier pénétrerait donc bien plus profondément dans la pente qu'il ne paraît.

Cet enfouissement du glacier dans la pente serait analogue à celui observé sous de petites taches de neige, à la formation de *niches de nivation* dans un terrain meuble. Ce n'en serait que la poursuite. Mes observations sur les niches de nivation en terrain meuble confirment celles de Demangeot (*) : « Il est remarquable que les niches de nivation installées en sol hétérogènes soient, l'été, le lieu de perte de filets d'eau superficiels : on peut donc admettre une évacuation du matériel fin par le fond de l'entonnoir, d'où une tendance à l'affaissement à cet endroit ». J'ajouterai que le sol de la niche devient, par disparition des fines, une couche de pierres (cf. fig. 5 prise dans le Cajon Barriga, à 4350 m).



Fig. 5 — Niche de nivation en terrain meuble.

5. EVOLUTION DES GLACIERS ENTERRÉS

Les glaciers enterrés, de même que les « névés permanents » ne satisfont pas à la définition traditionnelle du glacier : masse de glace d'origine nivale s'écoulant d'une zone d'alimentation vers une zone d'ablation. Tout le glacier est, suivant les années, zone d'accumulation ou zone d'ablation. Il suffit pour que cette masse de glace soit permanente, à l'échelle humaine, que le bilan annuel soit *en moyenne dans le temps* à peu près nul, et non en moyenne dans l'espace.

Toutefois par suite de son action morphologique le glacier enterré me semble avoir une évolution irréversible, même en supposant que le climat ne varie pas. Je résumerai ainsi son histoire :

1) A la suite d'une année exceptionnellement neigeuse, un petit névé subsiste à basse altitude, recouvert l'été d'éboulis.

2) Ce névé, rapidement transformé en glace, se creuse une niche, si bien que le relief extérieur de la pente ne varie pas. La présence de glace diminuant l'ablation, de la neige subsiste chaque hiver, si bien que le glacier s'accroît en épaisseur. Au fur et à mesure l'affouillement du sol sous-jacent par l'eau de fonte approfondit la niche, et il s'ensevelit de plus en plus.

3) L'épaisseur devient suffisante pour que le glacier s'écoule le long de la pente, entraîné par son poids. Il détruit alors le bas de sa niche et en sort, et prend sa forme caractéristique d'une langue *en relief* sur la pente.

4) Cette position en relief diminue l'alimentation en neige. Le glacier meurt, et sa glace disparaît progressivement. Seule subsiste une langue de débris meubles.

5) Cette langue de débris se transforme en un petit *glacier rocheux* (qu'il vaudrait mieux appeler : *glacier d'éboulis*). Des sols striés s'y forment, c'est-à-dire des drains de cailloux dans lesquels les parties fines ont été emportées par l'eau de fonte ⁽¹⁾. Ces drains sont alignés dans le sens de la pente dans la partie haute, pentue. Mais vers la langue ils suivent la direction des bourrelets que le mouvement a formés. Le mouvement vers le bas se poursuit, à vitesse réduite.

6) Toutes les parties fines disparaissent du glacier rocheux. Il ne comporte plus alors qu'un chaos de blocs qui s'immobilise et peut, sous d'autres climats, être envahi par la végétation.

Il n'est pas dit que tous les glaciers rocheux proviennent d'un glacier enterré, mais seulement les petits glaciers rocheux qui naissent en pleine pente. Ce n'est proba-

blement pas le cas pour ceux plus étendus que l'on trouve au milieu d'un cirque glaciaire net. De même il serait faux de croire que les cirques glaciaires ont même origine que les niches de nivation en terrain meuble. En morphologie, pour chaque ordre de grandeur il faut trouver une explication différente.

6. ENTRAÎNEMENT ET LAMINAGE D'UN GLACIER ENTERRÉ PAR UN GLACIER SURIMPOSÉ

A gauche de la fig. 1 on aperçoit une langue glaciaire, débordement du Glacier Olivares Beta par dessus la crête. Cette langue, actuellement en voie de récession, comporte une partie inférieure extrêmement riche en débris, gigantesque moraine de fond capturée pendant l'avance du glacier. La description de cette glace chargée de débris, extrêmement laminée, présentant des plis, des plis couchés et des plissements du deuxième ordre, a été faite ailleurs ⁽⁸⁾.

La très grande épaisseur de cette couche, au bout d'un écoulement de quelques centaines de mètres seulement, ne peut s'expliquer qu'en admettant qu'il s'agit d'un ancien glacier enterré, existant dans la pente avant le débordement. Le glacier surimposé l'a entraîné et laminé.

Cette observation me semble avoir une très grande importance. Un grand nombre de «moraines de fond», couches de glaces extrêmement riches en débris et à structure fluïdale, qui apparaissent au front des glaciers et des Inlandsis doivent s'expliquer par des phénomènes analogues, et non par une hypothétique remontée à travers la glace des matériaux arrachés au lit. Nous rejoignons, en précisant le détail des phénomènes, les idées de Boyé ⁽⁹⁾ sur l'érosion glaciaire : l'érosion glaciaire est due aux alternances de récessions des glaciers, pendant lesquelles le terrain se gélive et s'ameublisse, et d'avances glaciaires, pendant lesquelles le matériau meuble est évacué. Il apparaît que ce matériau meuble peut être déjà extrêmement riche en glace, soit qu'il s'agisse d'un sol gelé, soit, bien plus rarement il est vrai, de glaciers enterrés.

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SOME NOTES ON SLAB AND NICHE GLACIERS, AND THE CHARACTERISTICS OF PROTO-CIRQUE HOLLOWS

J.-M. GROVE

SUMMARY

The characteristics and locations of small, oval «slab» glaciers and a large ravine suggest that the floor of a proto-cirque hollow must slope at such an angle that the surface gradient of an ice mass accumulated within it is sufficiently great, but not too great, for rotational flow to take place, if it is to evolve into a cirque.

RÉSUMÉ

Les caractéristiques et les localisations de glaciers petits et ovale, ainsi qu'un ravin profond, suggèrent que la base d'une cavité pro-cirquienne doit glisser d'un angle tel, que la surface inclinée d'une masse de glace accumulée à l'intérieur, soit suffisamment grande, mais non excessive, pour permettre un courant de rotation qui évoluera dans un cirque.

Detailed accounts have recently been published of two cirque glacier in Midt-Jotunheimen, Norway, (Grove 1960 a & b McCall 1960). One of these, Vesl-Skaut-breen, is a simple glacier, occupying a cirque which is also unusually lacking in complication. The other, Veslgjuv-breen, is a compound glacier, which has formed a more complex cirque. Lying in close proximity, in each case, there is another glacier of comparable size, which is, like the cirque glaciers, roughly oval in plan. Neither of these glaciers has formed a cirque, although both are active. This situation raises the whole question of the pre-requisites for cirque formation and it therefore seems worthwhile to note some of the significant features of these ice masses. It is suggested that they might suitably be termed «slab» glaciers.

The eastern flank of Galdehö (2,222 metres) rises gently from the debris-covered plateau of Juvflyi, which is truncated by the western side of Visdalen. It supports three principal accumulations, two glaciers and a large snowpatch (See Figure 1). In the centre Veslgjuv-breen lies deep in its cirque, Veslgjuvbotn, of which the basin of Juvvatnet is morphologically an integral part. The incision into Galdehö is deep, about 1,300 metres horizontally from the western side of the lake to the foot of the headwall, which rises more than 200 metres above the ice and probably more than 400 metres above the deepest part of the floor. To the north, on the north-eastern slopes of the mountain, there is a large permanent snowpatch. There is no obvious evidence that it occupies a very pronounced hollow, since ablation causes retreat of the edges but does not reveal signs of a large underlying depression. Lastly, to the south of Veslgjuv-breen, and separated from it only by a narrow rock ridge, an unnamed glacier, about one kilometre long and half a kilometre broad, sweeps smoothly down from a height of about 2,100 metres at the top of the ice field to about 1,840 metres where the tongue reaches Juvflyi. The ice cliff at the northern margin is grounded at the edge of the lake. The glacier is a gently sloping slab of ice, with only a few minor surface undulations and it has rather larger area than Veslgjuv-breen, with which it was, until very recently, confluent. Its relative lack of morphological influence is most striking.

The head of the glacier is notched into the flank of Galdehö, but no backwall is visible, so that during the winter months the glacier surface appears as a continuation of the sweep of the hill side above it. However at the tongue, a terminal moraine

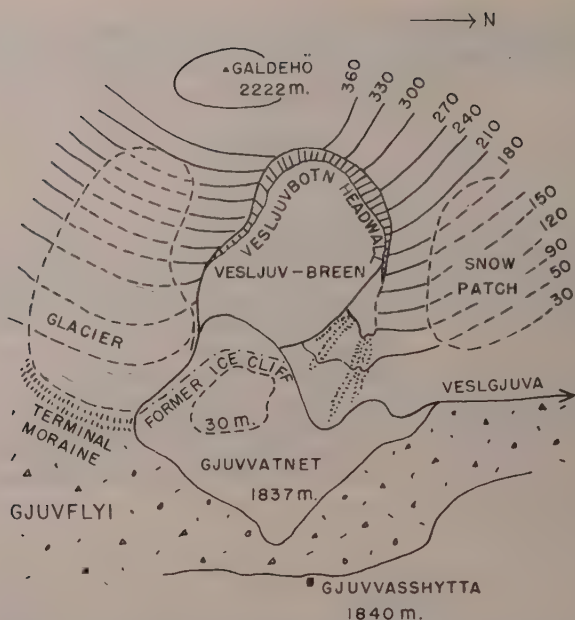


Fig. 1 — Sketch map of the eastern slopes of Galdehö—Approximate heights in metres above lake surface, at 30 metre intervals.

about 10 metres high, extends south from the southern end of Juvvatnet and eventually merges into low south-western laterals. Beyond this moraine irregular heaps of debris cover a zone about 100 metres wide. Most of this is fresh or moderately fresh in appearance, with little or no lichen cover. It is extremely variable in size, ranging from fine mud to boulders. As it is not possible for debris to fall onto the head of the glacier, the presence of such lateral and terminal moraines shows that some erosion must be taking place. The gentle slopes of the depression containing the slightly convex slab of the glacier can be seen at the southern margins. There are no sudden breaks of slope, but a smooth indentation of the slopes of Galdehö. The northern margins of the depression are concealed but during the last fifteen years recession of the ice cliffs in Juvvatnet has uncovered much of the northern face of the rock ridge separating it from Veslgjuvbotn. Towards the eastern end, the glacier still spills over the divide and enters the lake. In a zone about 100 metres up-glacier from the tongue caverns are often formed in the cliff, with resultant collapse of the ice front and parts of a crevassing area behind it. But this crevassing which probably marks the edge of the depression of Juvvatnet beneath the glacier, extends back, in intense form, for not more than about 150 metres from the cliff edge. The water is, in any case, shallow in this part of the lake and judging by the height of the terminal moraines, it is improbable that its floor has been radically remodelled by deposition from this glacier. The ice thus probably only enters the adjoining Veslgjuv-breen-Juvvatnet basin in the northern part of the tongue area and may be taken to be for the most part lying in its own far shallower depression. The question arises as to whether this hollow is a type of proto-cirque, which might, given time and suitable climatic conditions, develop to the mature form.

The glacier is certainly still active, as it is cut by crevasses, which vary in detailed position, but not in pattern, from year to year. One set, running parallel to the cliff edge



Fig. 2 — Galdehö and its glaciers from the north. Veslgjuvbotn and part of Velgjuv-breen can be seen on the right of the photograph, with the snow-covered Juvvatnet in the centre and the other unnamed glacier beyond them in the middle distance. Photo Widerøe's Flyveselskap A/s.

is associated with calving from the cliff in the ablation season. Its width is much less than that of the corresponding set in Veslgjuv-breen, where the basin form extends far beneath the ice. Another series runs north-south, parallel with the edge of the tongue, across parts of the lower third of the glacier. The crevasses are wider deeper and greater in number on the northern, cliff side of the glacier, where there is perhaps more change of floor gradient in the zone of approach to the subglacial extension of the lake basin. Other crevasses, running roughly north-south, in the marginal area where the ice is grounded above the southern precipice of Veslgjuv-botn are few in number and not strongly developed. The direction of flow appears, from the general orientation to be from west to east, that is directly down from the névé to the moraine. The surface gradient in this direction varies between about 11° and 14° . This is substantiated by the trends of the crevasses and the terminal moraines.

The glacier is built up of a series of accumulation layers, which outcrop on the surface as a regular series of bands. These run in gentle arcs, convex down-glacier, parallel with the general trends of the névé-lines and contours and at right angles to the assumed direction of flow. The curvature of the outcrops is no more than could be accounted for by the convexity of the surface and slight drag at the margins. While the plan of the outcrops is perfectly clear, there are often intrinsic difficulties involved in making detailed series of observations of the bands. The zone of frozen meltwater, usually found below the snowline in summer, (Schytt V. 1949) is abnormally wide on this glacier, and often extends almost to the tongue. This is probably because meltwater drains away over the entire ice surface. Normally, the ice surface is more irregular and much of the runoff is concentrated in streams and runnels, so that no wide and uniform ice crust obscures the lower part of a glacier. Moreover, stepping of the surface at the outcrops of ablation surfaces, a normal feature of most cirque glaciers in the region, is here uncommon and so the band pattern is all the more easily hidden by surface crusts and light falls of snow in summer. This lack of stepping

does not seem to be a direct function of slope. On the adjacent Veslgjuv-breen the most marked step formation is often found where the slope is less than 11° in a zone near the ice cliff. The glacier, as a whole, lies higher than Veslgjuv-breen, and in most years a much greater proportion of its area remains above the snowline. As the stepping is really a differential ablation effect, it may be that here even that part of the surface which is cleared of snow first, is not generally exposed for long enough for steps to develop, especially in view of the protection afforded by the formation of the ice crusts. It is also just possible that more fine debris from the surrounding bare ground is deposited on the surface of Veslgjuv-breen in summer, because of eddy effects set up by the cirque walls and that this is a further differentiating factor.

The result of these circumstances is that detailed or complete examination of the banding is usually difficult or impossible. However, a number of scattered observations of the inclination of the ablation surfaces, separating the accumulation layers have been made, and in every case they have been found to dip slightly down-glacier or to outcrop at the horizontal. The only exception has been in the immediate vicinity of the tongue. This ends in a small cliff about 1 metre high, in which many fine sheets of ice outcrop at about 45° to the surface, dipping inwards and giving a foliated appearance to the whole mass. Local shearing caused either by drag at the sole or by uprising where the flowing ice is blocked by the end moraine or a combination of the two might cause this foliation, which occurs also in the sole layers of cirque glaciers. (Grove 1960 *b*) Drag in the basal layers, to which it is confined, may be more important in forming this foliation, while overriding is probably responsible for the tilt. Some debris is included in the basal ice and the moraine disappears below the glacier in a mixture of rock fragments and ice. The only other complications of the accumulation layers were found in the ice cliff and its vicinity, where the outcrops are much disturbed and bear little relation to those exposed on the glacier. This is not unnatural, as the sections are marginal, atypical because of drag, disruption amongst the crevasses behind the face and drift deposition against the rock barrier.

The glacier is, then, built up of a series of accumulation layers, which by analogy with those of other glaciers, are probably annual. The form of the layers, originating in the névé, is probably lenticular, rather than wedge-shaped, because in the absence of a backwall accumulation tends to taper off in all directions from the centre of the névé-field. The fact that the accumulation layers are at or near the horizontal implies an absence of the rotational flow pattern which has been revealed in the contorted layers of the cirque-cutting glaciers.

It is most improbable that the glacier is filling a cirque so completely as to conceal its presence. Many cirques in the area are free, or almost free from ice and cirque headwalls are being revealed even in the central parts of the main massifs of Midt-Jotunheimen, where the present accumulation is greatest. Indeed it can be shown that the independent cirque glaciers are supported by the concentration of snow on them by drifting and avalanching, (Grove 1960 *a*). Therefore a cirque here completely filled by the ice to the top of its headwall would involve an anomaly in accumulation, which would be extraordinarily difficult to explain on climatic grounds. The glacier is contained in a hollow, which is probably not very deep, and which is being slowly enlarged as debris resulting from current erosion is carried away in the basal layers of the ice. There seems little reason for regarding this hollow as a proto-cirque. In order for cirque development to occur, a transformation of the flow pattern would have to take place. This could only be the result of an alteration of the accumulation pattern, which would be very difficult to envisage in the given topographical circumstances.

This glacier is not the only of its type in Midt-Jotunheimen. An example of a rather comparable form is found in the glacier which lies directly north of the summit of Spiterhø, to the east of Visdalen. This glacier is within a kilometre of Vest-Skaun

green, and is the larger of the two in area. The surface gradient is less than that of the Galdhøf Glacier and the ice mass is again built up of a series of horizontal accumulation layers. In this case there is a low backwall. Again it appears that the glacier lies in a broad, shallow depression, which is currently being enlarged, judging by the fresh moraine at the tongue. While the layers are wedge-shaped, it seems that the gradient is not sufficient for rotational flow to develop.

The view is frequently held that most cirques originate in valley heads. Certainly such a site must often have provided a suitable funnel-shaped collecting ground for snow, in which the steep, pre-existing head would favour rapid flow of the ice, as soon as sufficient thickness was attained. The importance of the part played by nivation in the initiation of cirques may be accepted. But snow patches may originate in hollows of many types, not only in valley-heads and gullies and they must tend to emphasise the depressions in which they lie. Indeed it has been suggested that nivation may be sufficient to convert shallow dimples on gentle slopes into infantile niches, which may later progress into true cirques. (Bowman I. 1916, Russel R.J. 1933, Lewis W.V. 1939, McCabe L.H. 1939). It may well be that a minority of cirques do originate in hollows formed largely or completely by nivation. Lewis (1939) working in Iceland, particularly associated circular snowpatches lying in slight concavities with potential cirque formation. But consideration of the characteristics of the two glaciers already described suggest that certain proportions are required if a nivation hollow is to be developed into a normal cirque. If it is too shallow or has too low a gradient, it may develop into a hollow which has only some of the characteristics of a cirque. It is such features that are being formed by the glaciers which have been described.

Other writers have emphasised that importance of ravines of various types (Richter E. 1900, Klebelsberg R.N. 1935) or landslip scars, for the beginnings of a cirque formation. A possible proto-cirque form might be that of the rasskars described by H.W. Ahlmann (1919) on the upper sides of Lyster Fjord, and according to him formed largely by stone slides or rockfalls, localised by fractures and zones of weakness, near the junction between the fjord-sides and the plateau surface above, and occurring rather late in the period of glaciation. Snow patches tend to lie in rasskars once formed, and no doubt to enlarge them further. Another possible sequence from steep water-eroded gullies, later modified to a funnel shape by nivation, through to a cirque form has been suggested by Groom (1959) in connection with a study of steeply hanging niche glaciers in Bünsowland, Vestspitsbergen. Here the influence of longitudinal rather than circular snow patches is involved and the niche glaciers are considered to be «glacier types genetically related to cirque glaciers, forming an early stage in corrie development». The exact nature of the early cirque form must be dependant both upon the local rock structure and upon the particular group of intermediary processes at work.

It is not difficult to find other possible intermediate ice forms in the sequence from longitudinal snowpatches to cirque glaciers. The peak, Styggehöe, on the west side of upper Visdalen, in the Galdhøpiggen massif of Midt-Jotunheimen, is gouged out by an extremely well marked hollow, in which there is a small glacier. The backwall and sides are very steep and the hollow extends from about 2,000 metres at the head, down to 1,400 metres. The whole floor appears to slant outwards. The glacier has well marked bands, outcropping across the surface at right angles to the directions of maximum slope. The hollow in which it lies is probably transitional between a large gully and a cirque, having some of the characteristics of each. It seems possible that it originated in a snowpatch occupying a steep ravine and that it could be the forerunner of a «crater» cirque the type which lies centrally in a mountain mass, such as that described by Cotton (1942, p. 175) in Norangerdal, or in that of Stolbtind in Hadselø, Vesteraalen described by Ahlmann (1919) p. 225.

It seems reasonable to accept that various types of gully and ravine have provided the conditions necessary for the early stages of enlargement and modification by snowpatches and niche glaciers, which may lead eventually to cirque formation. But, as with the sequence from circular nivation hollow to cirque, it seems that certain proportions may be necessary in the original depression and that, in this case, a limit may occur because nivation in a gully of more than a certain steepness may cause development of a large ravine rather than a cirque. An interesting example, which lends some support to this view, has been noted in the mountains of Ross-shire in Scotland. (See Figure 3).

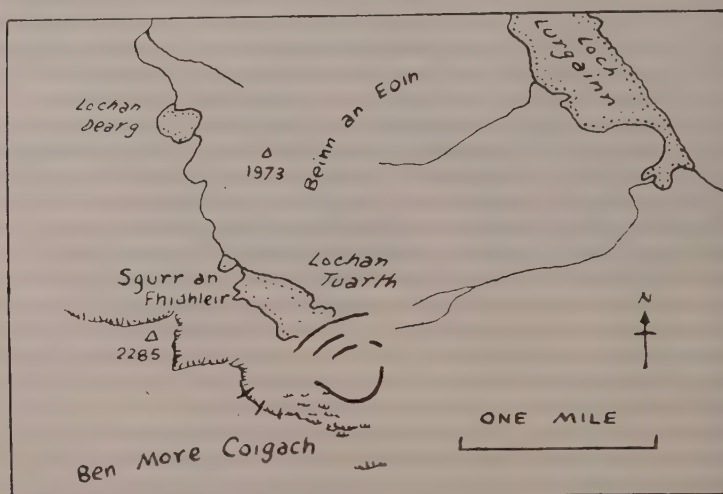


Fig. 3 — Sketch map to show position of cirque and ravines (Near Sgurr an Fhithleir).

A small cirque and a ravine of similar size have been cut, entirely in Torridonian Sandstone, in the north-east facing slopes beneath Sgurr an Fhithleir and Ben More Coigach. Both cirque and ravine must belong to a late glacial stage because they are incised into the southern wall of the arcuate glacial through-valley, which now drains north-west into Loch Lurgainn and separates the peak, Beinn an Eoin in the north from the massive Ben More Coigach in the south. The cirque, which is of normal simple type faces north and is fronted by a series of three concentric moraines, the outermost of which forms the watershed between north-east and north-west flowing streams on the valley floor. To the west, there is a group of steep, V-shaped ravines. Most of them are small features, but the ravine which lies immediately south of the subsidiary peak Sgurr an Fhithleir is comparable in scale with the cirque. The rockwall in which it is incised is, however, steeper than the one into which the cirque is cut. Both ravine and cirque must almost certainly have originated in gully formation and subsequent nivation and enlargement in each case was presumably much assisted by the cuboidal structure of the rock. Erosion on the steepest slopes has led to the cutting of ravines, while on the gentler part of the same wall a cirque has been formed.

From this it seems probable that some, but not all, snowpatch hollows are potential cirques, and that proto-cirques do not invariably form in valley heads. For example on glacially steepened slopes, snowpatch erosion is particularly likely to be assisted or localised by slide and creep effects. It seems that the floor of a hollow

must slope at such an angle that the surface gradient of an ice mass accumulated within it is sufficiently great for rotational flow to take place, if it is to evolve into a cirque. If the slope of the flow is too low, then the hollow may be enlarged by flowing ice, but will retain the same form as in the case of the hollows containing the «slab» glaciers described earlier. If the slope of the floor is too great, then an avalanching snow gully will result, leading to features like the ravines of Scurr an Fhìdhleir. The critical gradient necessary for rotational flow probably varies with the thickness of the ice. This may depend, in any case, upon local accumulation factors and speed of flow, as well as reflecting the initial depths of the hollow. Further information about the relations between ice thickness, surface gradient and inclination of ice structures would be most informative and before this is forthcoming no quantitative assessment of the situation can be given.

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ENERGY BALANCE DURING THE SNOW MELT PERIOD AT AN OTTAWA SITE

L.W. GOLD* and G.P. WILLIAMS*

SUMMARY

Observations were made on the energy balance at the surface of a melting snow cover for a two-week period in the spring of 1959 at Ottawa, Canada. The net radiation was found to contribute over this period 2305 cal/cm², and convection 785 cal/cm². Of this heat, 2290 cal/cm² were used for evaporation and 800 cal/cm² for melting. The water equivalent of the snow cover lost was 13.4 cm and of this, 3.4 cm is estimated to have evaporated. The observations are discussed in relation to climatic factors and similar studies carried out in other regions.

RÉSUMÉ

On a effectué des relevés relatifs au bilan énergétique tel qu'on le trouve à la surface d'une couverture de neige en train de fondre. Les relevés ont été effectués durant une période de deux semaines, au printemps de 1959, à Ottawa, Canada. Au cours de cette période le rayonnement a donné 2305 cal/cm² et la convection 785 cal/cm². Toute cette chaleur a servi à l'évaporation qui a pris 2290 cal/cm² et à la fonte des neiges, laquelle a pris 800 cal/cm². On a estimé que l'équivalent en eau de la couverture fondue a été de 13.4 cm et que sur ce chiffre 3.4 cm d'eau s'est évaporée. On analyse les relevés en fonction de facteurs climatiques et d'études semblables effectuées dans d'autres régions.

In the spring of 1959, observations were made at a site in Ottawa, Canada (latitude 45° 24' N, longitude 75° 41' W, elevation 103 metres (339 feet) above mean sea level) on the energy exchange at the snow surface during two weeks of the snow melt period. The purpose of the observations was to estimate, for the weather conditions that occurred, the amount of energy associated with each component of the energy exchange and to determine the evaporative and convective heat transfer coefficients as defined by an empirical equation similar to that used in the Lake Hefner studies (1). The site chosen for the observations was a treeless level field, 300 metres by 91 metres (1000 feet by 300 feet), with low buildings located along three of its sides and open to the west.

1. ENERGY BALANCE EQUATION

The energy balance at the snow surface for the two-week period is given by:

$$(1) \quad Q_r + Q_s + Q_e + Q_c = 0$$

where Q_r = net radiation, cal/cm²

Q_s = snow melt component, cal/cm²

Q_e = evaporative component, cal/cm²

Q_c = convective component, cal/cm²

Q_r was obtained by direct measurement. A first estimate of Q_s was obtained from measurements of the change in the depth and average density of the snow cover. From this, the weight of snow per unit area remaining after each day

* Snow and Ice Section, Division of Building Research, National Research Council, Ottawa, Canada.

could be calculated. To calculate Q_e and Q_c , the following equations were assumed valid:

$$(2) \quad q_e = k_e U_2 (e_a - e_s)$$

$$(3) \quad q_c = k_c U_2 (T_a - T_s)$$

where q_e = evaporative heat transfer, cal/cm² hr

q_c = convective heat transfer, cal/cm² hr

U_2 = wind speed at 2-meter level, cm/hr

e_a = vapor pressure of the air, mmHg

e_s = vapor pressure at the snow surface, mmHg

T_a = temperature of the air measured in a Stevenson's screen, °C

T_s = temperature of the snow surface, °C

k_e = evaporative heat transfer coefficient

k_c = convective heat transfer coefficient.

It was assumed also that Bowens ratio ⁽²⁾

$$(4) \quad R = \frac{q_c}{q_e} = 0.46 \frac{P (T_a - T_s)}{1000 (e_a - e_s)}$$

was valid. P is the atmospheric pressure in millibars. The average atmospheric pressure at Ottawa is very close to 1000 millibars. Since pressure fluctuations during the observation period were less than 2 per cent of the atmospheric pressure, the pressure correction was neglected. Therefore, combining Equations (2), (3), and (4) and neglecting the pressure correction, gives

$$(5) \quad \frac{k_c}{k_e} = 0.46, \text{ mmHg/}^\circ\text{C}$$

The convective and evaporative heat loss for the two-week period is given by

$$Q_e = \sum q_e$$

$$Q_c = \sum q_c$$

Substituting in Equation (1) gives

$$(6) \quad Q_r + Q_s = -k_e \left\{ \sum U_2 (e_a - e_s) + 0.46 \sum U_2 (T_a - T_s) \right\}$$

2. INSTRUMENTATION

The net radiation was measured with a Schulze radiometer. This instrument was calibrated for short wave radiation by comparison with an Eppley pyroheliometer. A cone, whose inside surface was painted with a flat black paint and whose surface temperature could be accurately controlled, was used to calibrate the radiometer in the long wave region. Some difficulties were encountered in the calibration of the instrument and in its use during observation. It is estimated that Q_r is accurate to within ± 10 per cent.

The wind speed was measured with a cup anemometer mounted at the two metre level. The air temperature was measured and recorded continuously with a ventilated thermocouple mounted 120 cm above the ground in a Stevenson screen. The vapor pressure of the air was obtained from the hourly readings of the wet and dry bulb temperature taken by the Meteorological Service of Canada at Rockcliffe Airport, about 1600 meters (one mile) from the observation site.

The temperatures at various levels in the snow cover were measured and recorded every 20 minutes with thermocouples connected to a Leeds and Northrup recorder.

The snow surface temperature was obtained by plotting these temperatures against depth and extrapolating to the surface. One thermocouple was usually located at the surface or very close to it. In the morning, the temperature of the surface usually rose very quickly to 0°C which helped to relieve the difficulty of determining the snow surface temperature when the thermometer was exposed to short wave radiation. The vapor pressure at the snow surface was assumed to be equal to that in equilibrium with a plane ice surface at the same temperature as the snow surface.

3. ANALYSES OF OBSERVATIONS

The period over which the energy balance was obtained was from 0000 hours 17 March to 2400 hours 30 March. The temperatures, T_a and T_s were plotted against time on one graph and the vapor pressures e_a and e_s on a second. The area enclosed by the T_a , T_s and e_a , e_s curves during periods of time when the wind speed did not change by more than 44.7 cm/sec (1 mph) was obtained with a planimeter. This area was converted to °C-hr or mmHg-hr and the resulting value multiplied by the appropriate mean wind speed for the period. These values were then summed over the total observation period and yielded $U_2(T_a - T_s)$ and $U_2(e_a - e_s)$.

In Fig. 1, the accumulated net radiation and weight loss of the snow cover are plotted against time. As a first approximation, it was assumed that all the snow was melted and so a value for Q_s was obtained. Using the observed values for Q_r , Q_s , $U_2(T_a - T_s)$ and $U_2(e_a - e_s)$, k_e was calculated from Equation (6). Using this value for k_e , a first approximation to the heat used in evaporating snow was obtained. The weight of snow evaporated was then deducted from the total weight and a new value for Q_s found. A new value for k_e was calculated using Equation (6). This procedure was continued until a change of less than 1 per cent in the previous value of k_e was obtained using the corrected Q_s . Equation (5) was used to obtain k_c .

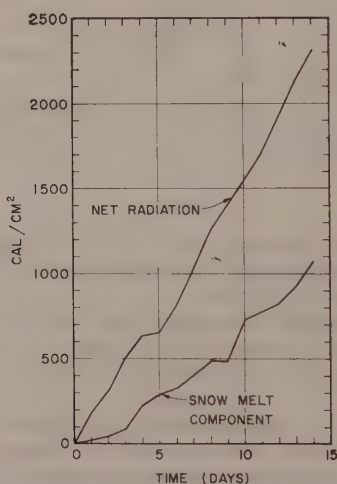


Fig. 1 — Integrated net radiation and snow melt component.

The observed value for k_e and k_c over the two-week period and the measured or calculated components of the energy balance are given in Table 1.

TABLE 1

*The components of the energy exchange measured or calculated for the 2-week period
0000 hours, 17 March to 2400 hours, 30 March.*

Q_r , net radiation	2305	cal/cm ²
Q_s , snow melt component	800	cal/cm ²
Q_e , evaporative component	2290	cal/cm ²
Q_c , convective component	785	cal/cm ²
$U_2 (e_a - e_s)$	-2.91×10^8	mmHg cm
$U_2 (T_a - T_s)$	2.18×10^8	°C cm
k_e	7.9×10^{-6}	cal/cm ³ mmHg
k_c	3.6×10^{-6}	cal/cm ³ °C

4. DISCUSSION

In carrying out a study such as this, the possible errors that may arise in the observations, assumptions, and associated calculations become increasingly apparent. It is of value to review briefly the possible errors that may occur.

With sufficient care, it should be possible to measure net radiation to within ± 5 per cent with instruments now available. In the present study, it is believed that the net radiation measurements were accurate to within ± 10 percent. The measurement and recording of the wind speed and the relative humidity and temperature of air inside a suitable screen a given distance above the snow surface presents no real difficulty. The present observations are considered to be accurate to ± 5 per cent.

The measurement and recording of the snow surface temperature presents a real problem. In the present study this was estimated by extrapolation of temperatures measured within the cover. Such a procedure is satisfactory at night but becomes questionable in the daytime when the temperature-measuring instruments are exposed to incoming short wave radiation. If the cover is actively melting, then the assumption that the surface is at 0°C, as was made in the present study, cannot be strictly valid as the heat required to melt the ice must pass through a film of water on the ice surface. Therefore, the air-water interface must be somewhat above 0°C. The authors are not aware of any studies that have been made that show if this effect would be significant. It is acknowledged that the estimate of the snow surface temperature, and therefore of the vapor pressure at the snow surface, are probably the largest source of error in the present study. Using the figures given in Table 1, the average difference between the shielded air temperature and the snow surface temperature is calculated to be 0.83°C. It is believed that this value is reasonable for the weather conditions that prevailed.

A number of theoretical equations for calculating the convective and evaporative components of the heat flow are available in the literature. Some of these require detailed and careful measurement of the temperature and vapor pressure gradient in the free air stream. Such measurements are often impractical for many field applications. Studies such as those done on Lake Hefner ⁽¹⁾ and by the co-operative Snow Investigations Group ⁽³⁾ have indicated which of the proposed formulae are probably most accurate, but they also show that from the practical point of view, equations such as (2), (3), and (4) are normally adequate. Priestley ⁽⁴⁾ discusses the different approaches for calculating the convective and evaporative heat flows and states that for a saturated surface, equations such as (2) are convenient for estimating evaporation and that the assumption contained in Equation (5) is probably satisfactory.

Measurement of the water equivalent of the snow cover does not present any real difficulty. The snow cover depth and average density vary with position and to sufficient observations must be made to ensure a statistically valid average. The information used in this study was the total weight of snow that has melted or evaporated over the two-week period and it is believed that the daily observations that were made determine this weight to ± 5 per cent.

The water equivalent of the snow that was lost was observed to be 13.4 cm and of this, 3.4 was calculated as evaporation loss and the remainder runoff. Attempts were made to measure the amount of evaporation using white aluminum pans and a technique described in a paper by Williams (⁵). One of the difficulties encountered was that during the daytime, the pan contained a mixture of snow and water and thus presented to the air a surface which differed from that of the surrounding snow cover. The observations were not continuous but those made show an evaporation loss less than that obtained by the preceding analysis. The observed evaporation would indicate that the calculated heat gain by convection is too high. If it is assumed that the convective heat gain is zero, then the calculated evaporation is 2.1 cm of water and this agrees favorably with what was observed. The value for the evaporative coefficient under these conditions is found to be 4.8×10^{-6} cal/cm² mmHg.

For a saturated surface, it is convenient to define an evaporative coefficient by the following equation from Priestley (⁴),

$$C_E = \frac{1.11 q_e T}{U_2(e_a - e_s)}$$

where T is the temperature in ° Kelvin, and

C_E is a dimensionless constant

Substituting for q_e from Equation (2) gives

$$C_E = 1.11 T k_e$$

Assuming $T = 265^\circ\text{K}$, C_E is found from the present observations to be 0.0023 for $k_e = 7.9 \times 10^{-6}$. There is some justification for assuming C_E to be equal to the drag coefficient used in momentum transfer calculations. Deacon found the drag coefficient over smooth snow on short grass to be 0.0012 and for a snow surface on natural prairie to be 0.0028 (⁴). Using $k_e = 4.8 \times 10^{-6}$ gives C_E a value of 0.0013. Since the surface of a melting snow cover is probably rougher than that of smooth snow on a short grass field, the higher values of C_E and k_e are indicated. A difference in surface roughness could account for the difference between the observed and calculated evaporative loss.

Accepting the results in Table 1 as giving a reasonable picture of the energy balance, it is of interest to consider some of the factors that affect this balance. Under the temperate conditions that existed at Ottawa during the period of observations, the snow cover disappeared when the weekly mean air temperature was about 0°C . Figures 2 and 3, showing the difference between the air and snow temperatures and the associated vapor pressure differences for two of the days of observations are typical for this period. For part of the time, the snow surface temperature was at its maximum of 0°C and the vapor pressure at its maximum of 4.58 mmHg. If the snow surface temperature and vapor pressure had not been limited in this way, as in the case for temperatures below 0°C , then the convective heat gain would have been less and the evaporation would have increased. Therefore, in regions of shallow snow cover, such as are found in the Canadian Arctic or Prairies, much of the cover can be lost by evaporation before the snow melt begins. Under such conditions, the heat lost by evaporation over a three or four day period would be very nearly equal to the net radiative gain less the change in heat content of the snow and ground.

On the other hand, if the snow cover can persist for some time after the snow melt has begun, for example as snow fields on north slopes, glaciers, or ice caps, then

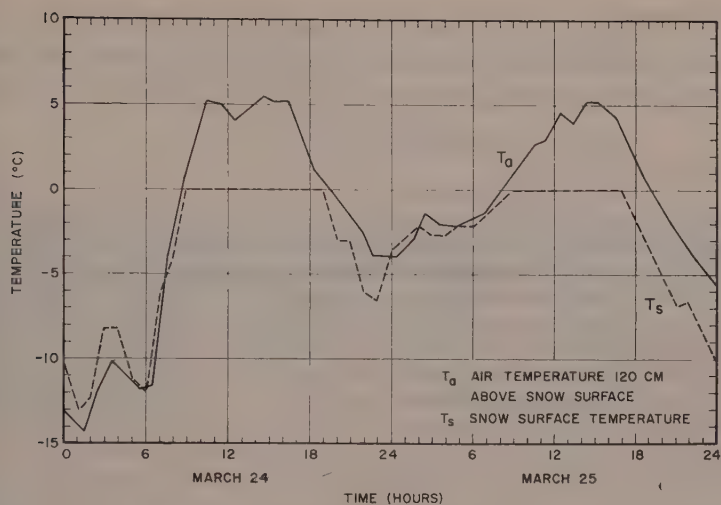


Fig. 2 — Free air stream and snow surface temperature.

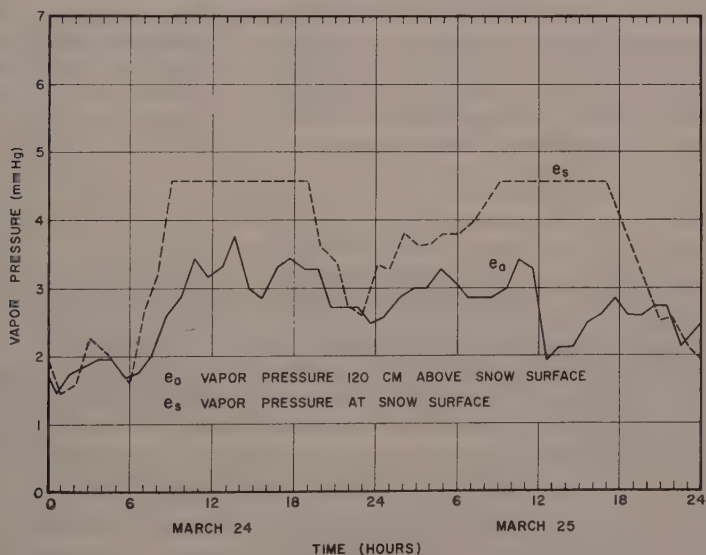


Fig. 3 — Free air stream and snow surface vapor pressure.

the air temperature may be persistently above 0°C and near saturation with the result that evaporation is reduced and even reversed and convection is increased. Under such conditions, both radiation and convection contribute to the snow melt component and the rates of melting can greatly exceed those observed in the present study. This is borne out by observations made by La Chapelle on the Blue Glacier in Washington State⁽⁶⁾, by Ward and Orvig on the Barnes Ice Cap in Baffin Land⁽⁷⁾, and by the Cooperative Snow Investigations Group⁽³⁾.

5. CONCLUSIONS

Under the conditions existing at the time of observation and for the assumptions made in the calculations, the evaporative heat transfer coefficient was found to be

$$7.9 \times 10^{-6} \text{ cal/cm}^2 \text{ mmHg}$$

and the convective coefficient

$$3.6 \times 10^{-6} \text{ cal/cm}^2 \text{ }^\circ\text{C}.$$

The net radiation and the evaporation were found to be the major components in the balance. The studies indicate that for snow surface temperatures below 0°C , the evaporative loss should be very nearly equal to the net radiative gain less the change in heat content of the snow and ground for periods of observation greater than three or four days. Snow surface temperature measurements are, at present, the greatest source of potential error in the energy balance observations. Further studies should be made to check the validity of the assumptions made and to determine, with reasonable accuracy, the constants and coefficients required for calculating the convective and evaporative heat exchange coefficients and their dependence on factors such as surface roughness.

6. ACKNOWLEDGEMENTS

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RADIATION TEMPERATURE OF THE SKY AND ABLATION OF ICE

Renzo ZANETTI

Istituto di Fisica Tecnica dell' Università di Padova

SUMMARY

The mean radiation temperature (MRT) of the sky can be found by employing the ROOSE formula (1938) suitably modified. Two frigorimeters Z are used which, at intervals of 10 or 5 minutes or less, give a dimension D which, placed in the suitable formula, rapidly supplies the MRT of the sky. This does not strictly follow either the temperature of the air or the diffused radiation.

This method will be first applied in a study of the Marmolada glacier.

1. GENERAL REMARKS

A considerable part of the thermal energy of bodies is exchanged through processes of radiation, in other words, of emitting and absorbing e.m. waves with a wave length greater than 0.4μ and infra-red waves. In these exchanges, which occur according to the formulas of reciprocal radiation, the differences of the fourth powers of the absolute temperatures come into consideration, taken two by two in all their possible combinations among the bodies.

The exchange of heat Q for a single radiation between a single body and all the others surrounding it can be considered to take place according to the equation

$$Q = ES(T^4 - T_{RM}^4)$$

where E represents the emissivity of the surface of the body with an area S , T represents its absolute temperature, and T_{RM} an absolute temperature called the *mean radiation temperature* (or MRT) of the system of all the other surrounding bodies. This dimension was introduced into the study of the biology to characterize the exchanges through radiation between the human body and the walls of a room, which are all supposed to have a temperature T_{RM} , different from that of the air.

In theory the phenomenon of the heat exchange of a body is the algebraic sum of all the innumerable exchanges with surrounding bodies of a very varying nature; position and temperature. One would therefore face practically insuperable difficulties in trying to make a direct calculation of the total heat exchanged through radiation.

Biologically the MRT can be determined indirectly by using Katathermometers and other instruments (HILL, DUFTON, VERNON, WINSLOW, etc.). In 1938 ROOSE⁽¹⁾ suggested the use of two frigorimeters, one with a darkened, the other with a silvered sphere. In 1947 GIOVANARDI and ZANETTI⁽²⁾ applied ROOSE's formula, using two frigorimeters Z in an original manner. In 1951 RICHARDS, STOLL and HARDY⁽³⁾, employing four small spheres of identical construction but differently coated, experimented the use of a *panradiometer* for the absolute measurement of the surrounding radiation.

But already in 1946 the writer had observed some anomalous figures of the MRT of the surrounding atmosphere when the apparatus was used in the open. Therefore on occasion of the 1st geophysical year, new instruments were prepared, not only to obtain an easy and rapid calculation of the MRT of the free surrounding space, and that of the total radiation of the sky, but more especially for the purpose of collecting a continuous record of MRT variations under all possible conditions, and studying their correlation with all other meteorological figures.

The first results were communicated in April 1959 at the Padua Academy of Sciences, Letters and Arts. Since then studies have been continued with the financial aid of the Italian C.N.R. (Centro Nazionale delle Ricerche).

2. MODIFICATIONS OF THE ROOSE FORMULA

The Roose formula can be written synthetically as follows:

$$Q_n - Q_r = (E_n - E_r) S (T_0^4 - T_{RM}^4)$$

where the Q 's represent the quantities of heat given off by the two spheres: darkened (n) and reflecting (r), E represents their corresponding emissivity, and T_0 the absolute temperature of the two spheres when they are in operation.

If this formula is used, anomalies are found in the figure of the T_{RM} , particularly when there is a clear sky. The anomalies are greatly reduced:

1) if the thermal flow radiated by each of the two spheres is divided into two parts; that of the hemisphere facing the sky, above the horizontal plane passing through the centre of the sphere, depends in fact on the MRT sought for, while that of the hemisphere below this plane depends on the radiation of the bodies on the surface of the ground, which have, to all effects, the same temperature as the air;

2) if one introduces a corrective figure representing the direct solar radiation acting on the sphere as if it were a flat disk with a surface $S/4$.

The formula (1) then becomes, for spheres maintained at 36.5°C .

$$(T_z : 100)^4 + (T_a : 100)^4 = 184 - D - 0.1 H. \quad (2)$$

where T_z represents the required absolute mean radiation temperature or temperature of the zenith (to distinguish it from the T_{RM}); T_a represents the absolute temperature of the air, D the difference of consumption per unit (in Kilocal, per square metre and per hour), and H the solar constant in the same units.

By darkening the spheres one obtains $H=0$, and from knowing the value of D , T_z can be calculated.

3. PLAN OF OBSERVATIONS

Until 1959, salutory observations were made with instruments merely designed to measure Q_n and Q_r . The main results are summarised in the communication to the Padua Academy. At present, with the aid of the C. N. R., new recording instruments are being constructed, some of which are already in operation.

The figures of D will be automatically obtained with a frequency of 5/10 minutes.

A second instrument will be installed on behalf of the S. A. D. E. of Venice in the neighbourhood of Marmolada glacier (Dolomites), to record the ablation of the glacier which feeds the Fedaia reservoir at an altitude of 2050 metres above sea level.

A special formula has been elaborated which will give the thermal flow per unit of surface of the glacier, on which ablation depends:

$$q_g = Q_r t_a : d - 0.9 D + 146.$$

where Q_r represents the cooling per unit that takes place in the reflecting sphere, d the difference between the temperature of the spheres and that t_a of the air, and D the difference of consumption supplied directly by the new instruments.

The formula will be especially applied during hours of diffused radiation with an overcast sky, and during the intermediate seasons when direct solar radiation is less intense and the data for solar radiation are more uncertain.

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THE ENERGY BALANCE DURING A 10 DAY ABLATION PERIOD ON THE GREENLAND ICE CAP

W. AMBACH

Physics Department of University of Innsbruck. Preliminary Communication

SUMMARY

Measurements of the energy balance of the melting ice surface were carried out within the «Expédition Glaciologique Internationale au Groenland» in summer 1959 in the area of ablation (Camp IV-EGIG) on the Greenland Ice Cap.

The amount of ablation was calculated during a period of 10 days including the individual contribution of the various factors. The difference between measured and calculated ablation was 3 per cent.

The incoming energy divides with about 90 per cent going into melting and 10 percent into heating the ice. The sensible heat, which is transferred by dynamic convection, is reduced by the latent heat since the conditions for evaporation are fulfilled more frequently than the conditions for condensation. Nevertheless, the amount of sublimed ice may be neglected.

These results are in good agreement with previous measurements of a 11 day ablation period which have been accepted for publication in the «Archiv für Meteorologie, Geophysik u. Bioklimatologie.»

1. INTRODUCTION

«Expédition Glaciologique Internationale au Groenland» (E. G. I. G.) is a joint expedition of Denmark, France, Germany, Switzerland, and Austria. It was founded in 1956 and is under the patronage of the Commission of Snow and Ice of the A. I. H. S. (Association Internationale d'Hydrologie Scientifique). Program and organizational details are published (Bul. A. I. H. S. no. 6, 1957). The entire expedition is subordinated to an international management committee (president: Dr. B. Fristrup, Denmark). The *Expéditions Polaires Françaises* were charged with the technical execution, their director, P. E. Victor, is also head of the expedition. The scientific results were published in the Danish periodical *Meddelelser om Grønland*.

The present investigation was carried out within the team «Glaciologie Côtière» (teamleader: A. Bauer) during the 1959 summer-campaign.

Austria's contribution to the E. G. I. G. was a study of the heat balance in the area of ablation on the Greenland Ice Cap (Camp IV-EGIG 49° 37'W, 69° 40'N, approximately 1000 m altitude).

2. CALCULATION OF THE COMPONENTS OF THE ENERGY BALANCE

The special position occupied by the ablation area of the Greenland Ice Cap in the studies of energy balance has been pointed out in an earlier paper⁽²⁾.

The positive energy surplus in the ablation area of the Greenland Ice Cap is consumed both for melting and for heating the ice. The heat which is conveyed by molecular conduction from upper ice layers into lower ice layers increases the temperature of the ice. This heat flow is fed by the energy surplus on the surface. The processes by which energy is exchanged on the surface are the following: radiation Q_R , dynamic convection Q_C , and precipitation Q_P . The heat balance equation furnishes the energy balance of the surface:

$$Q_R + Q_C + Q_P = Q_M + Q_H$$

where Q_M denotes the energy consumed for melting, and Q_H for heating the ice. All terms of this equation may have positive or negative signs. The energy exchanged by precipitation may be neglected in almost all cases.

2.1. Radiation measurements

For measuring the short-wave radiation two Moll-Gorczyński solarimeters (Kipp and Zonen, Delft) were used. A thermopile was directed upwards for measuring the incident energy; the thermopile of the second solarimeter was directed downwards for measuring the reflected short-wave energy. Furthermore the net incoming and outgoing radiation was measured by means of an R. Schulze net-radiometer (Lipolem instrument) with the incoming and outgoing energy being recorded separately. The calibrations in the short-wave spectral range were carried out in situ for all four thermopiles by using a Linke-Feussner actinometer (Kipp and Zonen, Delft).

The calibration in the long-wave range of the spectrum was made by means of a black body. Point recorders (Hartmann and Braun) were used for recording.

In determining the incident short-wave radiation energy from the recordings of the solarimeter the strong dependence of the calibration factor on the solar altitude was taken into account. The calibration curve for various solar altitudes between 42 degrees and 0 degrees has been published in (3). The calibration factor for diffuse isotropic radiation was determined according to a method given by G. H. Liljequist (7) for the thermopiles of the solarimeters and of the net-radiometer. The evaluation was made according to mean values per hour. For time intervals without direct sun the calibration value obtained for isotropic radiation was used. For hourly intervals with interrupted direct insolation the mean value could be approximated from the calibration values for direct insolation and isotropic insolation. The temperature correction for the radiation sensitivity of the thermopiles was applied to all hourly mean values. The value found by H. Hoinkes (6) for the same radiometers in the Antarctic was used as a factor of temperature correction. In this case the temperature measured by the thermocouple of the net-radiometer was inserted as instrument temperature for the solarimeter as well. In addition the change of the resistance of the current circuit due to different temperatures was taken into account. For calculating the short-wave radiation absorbed by the ice albedo values were used which were obtained by means of a portable solarimeter (Kipp and Zonen, Delft) at the sites of ablation measurement. The mean albedo was determined to be 43, 1% from 50 individual values. The value of long-wave radiation balance was estimated by means of an expression which H. Hoinkes and N. Untersteiner (5) deduced from measurements on an alpine glacier. W. Ambach (1) obtained very similar values. The expression used is:

$$A_i = 0,085 \left[1 + 1,4 \left(\frac{i}{10} \right)^2 \right] \text{ ly/min}$$

where A_i denotes the long-wave radiation balance and i the cloudiness in tenths.

For the purpose of calculating a daily mean value for the short-wave radiation balance the mean daily cloudiness was inserted into this expression.

2.2. Dynamic convection

By calculating the exchange coefficient from the logarithmic increase of wind velocity with height, the flow of sensible and latent heat can be calculated if the gradients of temperature and vapor pressure in the air layer near the ice are known. The simplifications and assumptions made earlier (3) were maintained. The values of air temperature and of vapor pressure were measured in a meteorological screen 187 cm above the surface of the ice. The mean value of air temperature for this period

was found to be 1,90°C and for vapor pressure 4,52 torr. From the wind profile the friction velocity was found to be 42,6 cm/sec and the exchange coefficient 2,16 g/cm sec. The roughness parameter was 0,4 cm.

In order to fulfill the assumptions it is necessary that the temperature of the ice surface be 0°.

The ten-day ablation period under discussion was chosen from this point of view. The period begins after the end of a cold wave and ends with the beginning of a short period with considerable night frost. During the ten-day ablation period between them positive temperatures were continuously recorded in the meteorological screen. Nevertheless, the surface may have been slightly frozen for short periods. This error, however, will probably not affect the results appreciably.

2.3. Heating of the ice

By measuring the ice temperature down to a depth of eight meters it is possible to determine the heat which is conveyed by molecular heat conductance into deeper lying ice layers. The results of the series of measurement II and III which were carried out directly at the station were used for the calculation. For the ten-day period the variations in the heat content of the ice differ only by 20 ly. For the purpose of calculating the energy transformation the mean value from both series of measurements were used. The influence exercised by the heat which flows upwards from depths of more than eight meters can be neglected for this estimation.

2.4. Ablation measurements

The ablation measurements were made at ten points over a radius of about 50 m. For calculating the energy transformation the mean value obtained from these ten ablation stakes was used.

It may be seen from the mean daily ablation value of 3,74 cm that the ablation period under discussion is characteristic of the entire ablation period of the summer of 1959.

3. RESULTS

In determining the energy balance on the surface it is necessary to determine all components separately. Only then can the heat balance equation be checked by calculating the ablation from the energy balance and comparing it with the measured value. The following table gives all energy values calculated for the period from July 12, to July 21, 1959, in a total of 10 days.

short-wave insolation: 5250 ly	energy gain	energy loss
absorbed short-wave radiation energy, albedo 43,1%	+ 2980 ly	
long-wave radiation balance		— 300 ly
sensible heat flow	+ 460 ly	
latent heat flow		— 100 ly
heat conduction into the ice		— 260 ly
sum	+ 3440 ly	— 660 ly
energy available for melting	+ 2780 ly \cong 34,8 g \cong 38,2 cm of ice	
energy consumption for evaporation	— 100 ly \cong 0,2 g \cong 0,2 cm of ice	
total ablation computed		38,4 cm of ice
total ablation measured		37,4 cm of ice
difference between measured and computed ablation		3%

These results are in good agreement with the values of an eleven-day ablation period from June 27, to July 7, 1959⁽⁶⁾ which have already been published. The relatively great contribution of radiation to the energy balance on the surface of the ice and the relatively small contribution of the energy supplied by dynamical convection was again observed since the sensible heat flow generally supplies energy to the surface whereas the latent heat flow on the average consumes energy by evaporation. From the positive energy balance of 3040 ly, 8,5% is consumed for heating and 91,5% for melting the ice. Although the condition for evaporation is fulfilled more frequently than the condition for condensation, the amount of the ice evaporated is very small and it can almost be neglected in ablation analyses. In the energy balance however, the influence exercised by the evaporation process must be taken into account because of the high sublimation heat of the ice.

These results cannot be directly compared to the values of energy transformation obtained by *M. Diamond and R.W. Gerdel* in the accumulation area about 200 miles east of Thule Air Base at an altitude of 6900 ft. The energy transformation in the accumulation area strongly differs from that in the ablation area.

The author expresses his gratitude to the *Österreichische Akademie der Wissenschaften* (Austrian Academy of Sciences) for their financial support without which this task could not have been accomplished. Grateful acknowledgment is also made of the financial support rendered by the *Expéditions Polaires Françaises* (Missions *Paul Emile Victor*) in the evaluation of results, as well as of numerous helpful discussions with *Prof. H. Hoinkes*, Head of the *Institute of Meteorology and Geophysics of Innsbruck University*, who also provided the radiation instruments used during this expedition.

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RADIATION AND THERMAL FACTORS IN THE ABLATION OF GLACIERS ON THE SOUTHERN SLOPE OF ELBRUS

A. VOLOSHINA (U.S.S.R.)

SUMMARY

1. The albedo of glaciers, situated at various heights, is connected with soulage, which as a rule decreases with height. Moreover the position of the glacier and the meteorological factors (precipitation and wind) are of great importance. The average dimensions of the albedo of glacial tongues (2400-3400 m) is 0.2 to 0.3 and of high plateaux (3700 m) 0.5 to 0.6; the maximum dimensions recorded on the saddle of Elbrus (5300 m) are 0.7 to 0.8.

2. Maximum degree of direct solar radiation on a perpendicular surface at a height of 2.4 km. amount to 1.54 cal/cm^2 per min., and at height of 5.3 km to 1.74 cal/cm^2 per min. Transparence of the atmosphere at over 5 km. is almost ideal.

3. Differences in total radiation measured at different heights are insignificant (not more than hundred of a calory).

4. Differences in the dimensions of radiation balance at different heights are determined by the albedo. The largest quantity of radiation heat is received by the glacial tongues (on clear days $400\text{-}500 \text{ cal/cm}^2$ per day). Moreover, the negative dimension of the balance is not more than $60\text{-}70 \text{ cal.}$, i. e. nearly 15 % of the daily total. On the firm plateaux the daily amounts of radiation heat come to only 250 cal/cm^2 per day; the net loss of heat is nearly 60 cal. , i. e. 25 % of the total amount. At heights of over 5 km the radiation balance is near to zero.

5. The wind regime, temperature and humidity play a definite part in the ablation of glaciers. The inverse distribution of temperature in the ground stratum is maintained at great heights, which indicates the penetration of warmth from the air. With positive temperatures the role of the radiation heat in ablation shows a sharp increase. On the glaciers of the southern slope of Elbrus in summer the positive temperatures during the day are observed up to a height of 4.0-4.2 km.; day and night, up to 3.2-3.5 km.

6. The distribution of moisture in the ground stratum shows that in valleys condensation of moisture predominates on glacial tongues, whereas on open elevated plateaux evaporation can be observed all round the clock (in the nature of 1-2 mm of water per 24 hours).

7. In the summer period the upper three-metre layer of a glacier up to heights of 4.0-4.2 km is characterized by isothermal stratification and has temperatures close to zero; this indicates the absence of warm currents in this layer in the daytime.

ENERGY EXCHANGE MEASUREMENTS ON THE BLUE GLACIER, WASHINGTON

E. LA CHAPELLE

Dept. of Meteorology & Climatology
University of Washington

SUMMARY

The Blue Glacier is a temperate alpine glacier in a strongly maritime climate. Heavy winter snowfalls establish a large energy deficit by ice mass accumulation. Mild air temperatures and thick snow cover insulation severely limit penetration of winter chill in the glacier; cooling is observed in a surface firn layer only a few meters thick, and represents about two per cent of the total winter energy deficit. Clear weather is rare in winter and radiation exchange at the snow surface is negligible for long periods.

At the low elevations of this glacier (1300-2250 meters) summer ablation is heavy. Heat exchange evaluations by the energy balance method during fair weather have shown solar radiation responsible for 69 per cent of the heat supply, while eddy conduction and condensation furnished 25 per cent and 6 per cent respectively. Sixty four per cent of the available heat is used for snow melt, the balance being lost to radiation cooling (28 per cent) and evaporation (8 per cent).

Sensible and latent heat transfer between snow and air have been calculated by the Bowen Ratio method and from aerodynamic relations. These values are in poor agreement with the energy balance values above. The discrepancy is attributed to difficulties of observing and interpreting temperature, vapor pressure and velocity profiles in air which is incompletely modified by flow over the glacier surface. Daytime air temperature profiles with maxima near the surface are common, and highly irregular wind profiles are often introduced by shallow drainage winds.

Investigation has shown that sub-surface snow melt due to transmitted solar radiation occurs to a depth of 15 to 20 cm. Special techniques have been developed to include measurement of this melt in snow ablation observations. Other radiation studies both theoretical and applied, demonstrated that light diffusion in the 5-cm surface snow layer has an important influence on the surface albedo.

1. LOCATION

The Blue Glacier is located on the northern slopes of the Mt. Olympus massif in western Washington State, USA. Mt. Olympus, 2415 meters above mean sea level, is the highest summit of a group of peaks forming the central highland of the Olympic Peninsula. (Fig. 1) The glacier originates in two accumulation basins reaching up to 2375 meters elevation, from which two separate ice falls descend to form a single valley ice stream at 1750 meters. This ice stream flows between mountain walls to the bifurcated terminus at 1275 meters. The west accumulation basin is shared with the adjacent Black Glacier, the surface ice divide being indicated in Fig. 2. The micro-meteorological station is located along this divide on a nearly level snow plateau, the Snowdome, while the research station buildings and climatological observatory are situated on a nearby rock ridge.

2. CLIMATE

Mt. Olympus rises well above the surrounding peaks, and the Blue Glacier thus occupies an exposed position above an area of long ridges and deep river valleys which are heavily blanketed by coniferous forests.

The Pacific Ocean, 55 airline kilometers distant from the glacier, imposes a true maritime climate on the western Olympic Peninsula. Annual precipitation varies

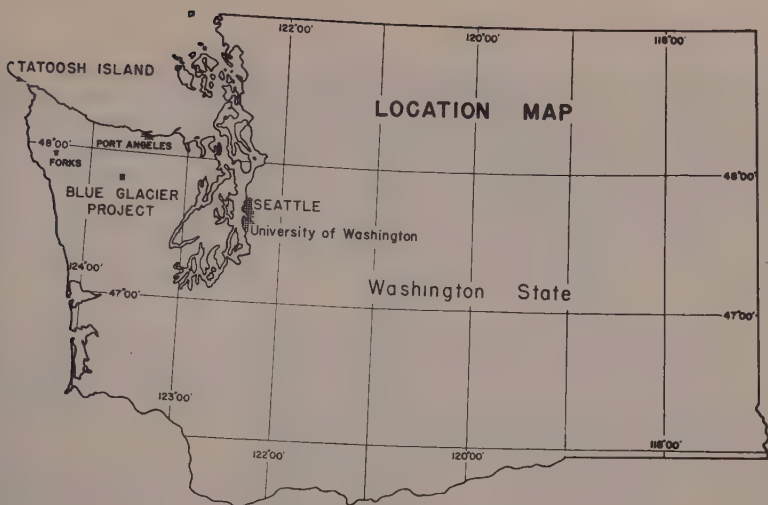


Fig. 1 — Location map (line drawing).



Fig. 2 — Blue Glacier map (line drawing).

from 400 to 500 cm a year in the center of this area, the wettest region in the conterminous United States. By far the larger part of this precipitation falls between September and June, and above 2000 meters elevation most of it falls as snow. During the IGY winter 1957/58, observers on the glacier recorded a sustained average precipitation rate of 2.5 cm per day from mid-December until early March. Clear weather during this part of the year is rare, the glacier often being blanketed by clouds and snow storms for days or even weeks at a time. Above about 2250 meters the persistent clouds, frequently driven by high winds, deposit thick accretions of rime and rime-cemented snow which may attain a thickness of 50 to 100 cm even on vertical rock walls.

The summers are frequently fair and dry, the months of July, August and September being the period of minimum precipitation. May and June are climatically the most variable months which may be given over to either predominant accumulation or ablation according to prevailing weather patterns in any given year. Midsummer ablation of snow averages around 2.5 cm of water per day.

After a prolonged period of net annual mass loss, the Blue Glacier apparently experienced a series of net mass accumulation years between 1948 and 1956, though quantitative data are lacking. This resulted in cessation of the terminal retreat which had occurred for many years. The net mass budget was moderately negative in 1957, strongly negative in 1958, and near equilibrium in 1959.

3. WINTER ENERGY EXCHANGE

The dominant factor in winter energy exchange is the formation of a large latent energy deficit through the accumulation of ice mass in the form of a deep winter snow cover. Snow depth in late winter normally attains a depth of 5 to 6 meters on the Snowdome, and more in the cirque. Total annual ice mass accumulation averages 350 gm/cm² for the glacier as a whole.

Energy available from solar radiation during the midwinter months is negligible. The predominantly cloudy and stormy weather reduce incident radiation at the glacier surface to a very low level for long periods of time, while albedo of the fresh snow under these conditions of diffuse illumination is generally 0.90 or higher. Total solar radiation absorbed by the snow surface during December, January and February, totalled approximately 1400 cal/cm².

The same conditions of fog and storm impose a uniform temperature regime at the snow surface and on the adjacent air and atmospheric moisture. Long wave radiation exchange thus is also negligible under these conditions. Frequent periods of riming and precipitation which render inoperative the net radiometers preclude a quantitative estimate of this small heat exchange.

The only significant winter periods of radiant energy exchange are the two or three days of clear weather encountered each month, at which time appreciable nocturnal radiation cooling of the snow surface takes place. Effects of this cooling are confined to a shallow surface layer of snow.

Winter snowfalls on the upper glacier normally occur at prevailing temperatures between 0° and -5°C. Snowfalls close to the freezing point are common. The relatively warm character of the accumulated snow cover, plus the insulating effect of the thickness quickly attained by this cover in early winter, prevent any strong cooling to the underlying firn or ice. Occasional winter rainfalls further hinder the formation of strong temperature gradients in the snow cover. (Fig. 3). At the Snowdome micrometeorological station, subfreezing temperatures in the winter 1957/58 penetrated a maximum of 4 meters into the firn. Except for short periods in November when the snow cover was shallow, firn temperatures never fell below -2°C. The energy deficit engendered

DISTRIBUTION OF SUB-FREEZING TEMPERATURES IN SNOW AND FIRN

Snowdome study area

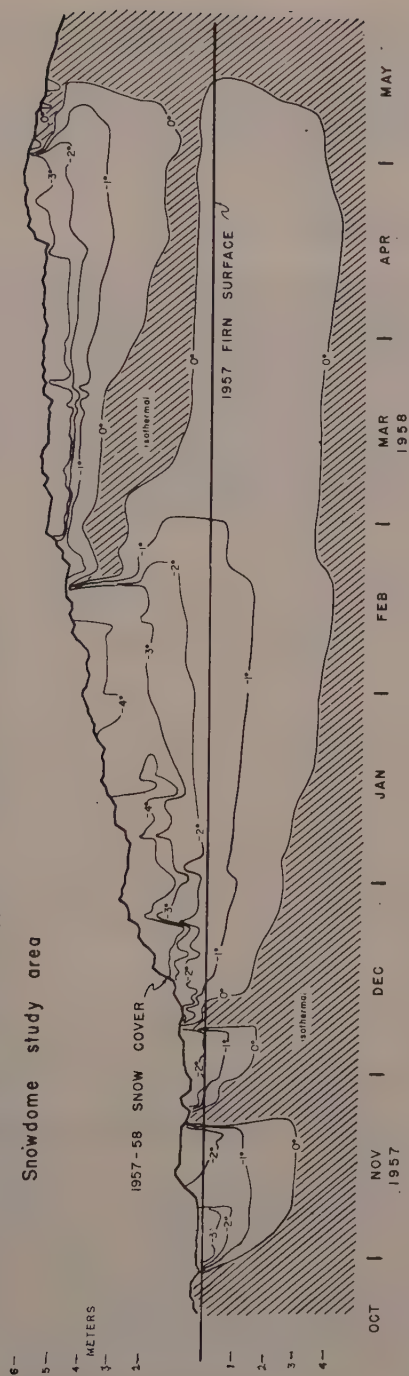


Fig. 3 — Annual trend of isotherms in snow and firn (graph).

by cooling of ice or firn and addition of subfreezing snow in 1957/58 is estimated to be about 2% of the total; the rest may be attributed to the latent deficit introduced by accumulated ice mass.

These sub-freezing temperatures are mostly dissipated by refreezing of percolating meltwater in the spring, rather than by conduction. Even the small temperature energy deficit thus is converted ultimately to a latent energy deficit.

4. SUMMER ENERGY EXCHANGE

After the winter chill has been dissipated, a temperate glacier serves as a heat sink which absorbs latent heat of fusion from its environment without change in temperature. On the Blue Glacier such a state persists steadily from May or June until about mid-September. Ablation in the fall is irregular, often being interrupted by early snow storms or periods of sub-freezing air temperature, although some melting may persist as late as November. This glacier absorbed energy at a maximum rate near the end of July in 1958, resulting in a maximum meltwater discharge at this time (Fig. 4). Such a curve for total glacier ablation of course depends on the geometry of the glacier, the rate of snow line retreat and relative areas exposed of snow and ice, in addition to variations in the external energy supply.

DAILY MELT RATE FOR TOTAL GLACIER SURFACE Blue Glacier 1958 melt season

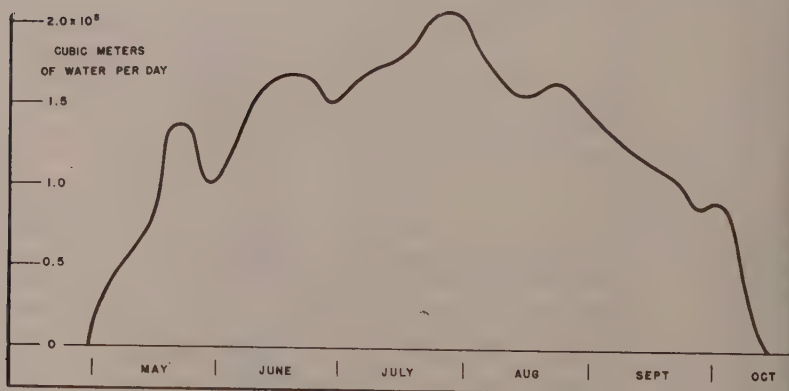


Fig. 4 — Total glacier ablation rates (Graph).

During mid-summer months a large high pressure area normally dominates the North Pacific, with frequent long periods of fair weather in the western part of Washington State. Under such circumstances the predominant source of energy for ice melt on the Blue Glacier is solar radiation. In the fall this is no longer true, for available heat from the sun decreases and the surface albedo rises from occasional fresh snow deposits. Condensation and sensible heat transfer from the air during fall storms then become the most important heat sources.

Micrometeorological studies over a melting snow surface were conducted on the Snowdome during the summers of 1958 and 1959. Instrumentation included solarimeters to measure incoming and reflected solar radiation; net and total radiometers used to determine total radiant energy flux at the snow surface; highly sensitive anemometers to obtain the vertical wind profile; and thermocouple elements for measuring the vertical air temperature gradient. Vapor exchange measurements initially were made with a lysimeter and intermittent psychrometric observations. Later these were supplemented with vapor pressure profiles determined by wet and dry bulb thermocouples. All the radiation instruments, the thermocouples, and the sensitive anemometers could record continuously.

Observations were obtained for a 37-day period of almost continuously fair weather in 1958. Results have been computed for this period, taking eddy conduction of sensible heat as the residual in the equation of energy exchange. Solar radiation furnished 69 per cent of the heat supply, while eddy conduction and condensation supplied 25 per cent and 6 per cent, respectively. Snow melt absorbed 64 per cent of this heat, evaporation 8 per cent, and the balance, 28 per cent, was lost to radiation cooling.

Studies in 1959 examined in more detail the character of sensible and latent heat transfer between snow and air at this location. Values of vapor flux at the snow surface as computed from the wind and vapor pressure profiles have to date agreed rather poorly with these values obtained by direct measurement of the energy balance method. This is largely attributed to the peculiar character of the observed wind and temperature profiles at the study site. A common feature here is a shallow drainage wind whose maximum velocity, which seldom exceeds 1.5 meters per second, occurs 1 to 2 meters above the snow surface. During fair weather this drainage wind interacts in a fitful fashion with the light prevailing wind, resulting in highly irregular profiles.

Even when light prevailing winds are dominant, air temperature profiles which show a maximum near the snow surface are frequently observed. (Fig. 5) These have a definite diurnal character, appearing only during the daylight hours in fair weather, and have been observed with a sufficient variety of sensing elements to suggest that they do not result from instrument errors. Though the origin of such an air temperature distribution above the melting snow is obscure, the local warming of surface air layers by sun-illuminated rocks at the glacier margins is considered a possible mechanism. Such profiles obviously complicate the computation of sensible heat transfer by means of the usual atmospheric turbulence theories. The origin of this temperature anomaly is being investigated during the summer of 1960 by means of two-dimensional instrumentation designed to examine the down-wind modification of air as it passes over the glacier surface.

During fair weather the vapor exchange exhibits a marked diurnal fluctuation in both sign and amplitude. Condensation predominates during the day, especially in the afternoon, while evaporation is the rule at night (Fig. 6). The condensation period coincides with a rise in the free air dew point as noted at the climatological station, and with the afternoon formation of cumulus clouds over the Olympic Mountains. In such conditions the glacier appears to be dominated at night by a prevailing dry air mass whose dew point commonly lies below 0°C , while solar heating of the surrounding forested valleys results in convective uplift of more moist air during the day. Martinelli (1), working on high altitude snow patches of the Colorado Front Range, observed dew point variations with a similar diurnal character but of opposite sign, the condensation period occurring at night.

Because solar radiation is the primary source of heat for summer ablation, its effects on the surface snow layer has received special attention. As distinguished from apparent surface wastage or ablation, the change of phase, or ice melt, in the snow is an important element in the equation of energy exchange. Transmitted solar radiation is responsible for measureable amounts of melting as deep as 20 cm below the summer

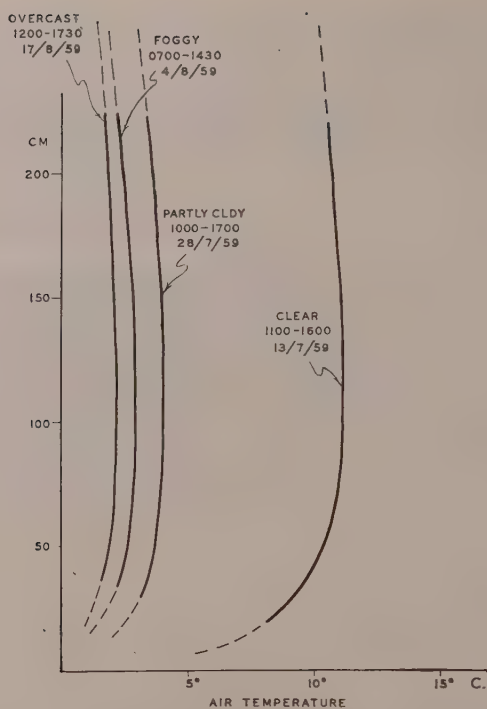


Fig. 5 — Anomalous air temperature profile (graph).



Fig. 6 — Diurnal variation of vapor pressure gradient (graph).

snow surface. This melt is not necessarily reflected by a change in surface level. This discrepancy between wastage and ice melt must be taken into account if the fair weather energy balance is to be correctly assessed. A technique has been devised whereby total ice melt over a given interval can be measured by determining graphically the area enclosed between two successive curves showing the distribution of ice density with depth. Such a plot is shown in Fig. 7. The necessary ice density values are obtained by measuring bulk density with small sample tubes and correcting for the free water content extracted with a hand-operated centrifuge.

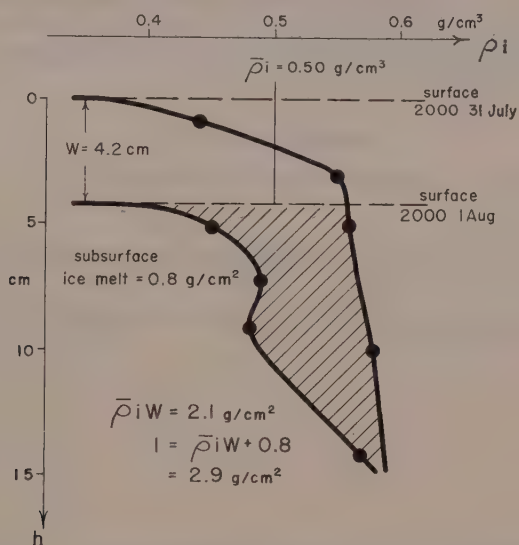


Fig. 7 — Ice density profiles in surface snow layer (graph).

The distribution of solar radiation within the snow cover is difficult to measure directly, for the introduction of sensing elements within the snow immediately alters the local radiation environment. A more elegant approach would be to deduce this distribution from observations made with instruments external to the snow cover. Using diffusion equations, Dr. J. C. Giddings of the University of Utah has predicted the variation of surface albedo with thickness of a snow slab whose undersurface is a perfectly absorbing (black) medium. This prediction has been tested on the glacier, using coarse-grained summer snow and photoelectric cells equipped with narrow-band filters. Checks have also been made under laboratory conditions using a monochromatic light source and fine-grained, cold winter snow which has been homogenized by disaggregation. Only partial confirmation has been obtained to date. Some practical results have been a quantitative demonstration of the fact that visible light reflected from the snow surface originates at depths up to 5 cm below the surface, and that the apparent monochromatic absorptivity of a snow slab varies with slab thickness.

5. CONCLUSION

In summary, the Blue Glacier has been characterized by the studies to date as a temperate alpine glacier situated in a strongly maritime environment, where the winter energy deficit is due almost entirely to accumulated ice mass, and where the principal

source of heat for summer melting comes from insolation. Sensible heat transfer from the atmosphere plays a welldefined but secondary role in the ablation.

Vapor exchange participates measurably in the energy exchange, and undergoes a distinct diurnal fluctuation in fair weather.

Irregular shallow drainage winds and anomalous air temperature gradients have combined to render difficult the computation of sensible and latent heat exchange between air and snow. The sensible heat exchange appears to be obtained most readily as the residual in the energy balance.

Sub-surface snow melt plays a significant part in the daily energy balance. Special techniques of ablation measurement have been developed to include this quantity in the measured phase change. Distribution of solar energy beneath the snow surface has been investigated by indirect methods.

REFERENCES CITED

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INFLUENCE DU CLIMAT SUR LES GLACIERS

RESPONSE OF GLACIERS TO CLIMATE

TEMPERATURE OF ICE IN THE LOWER PARTS OF THE TUYUKSU GLACIERS

E.N. VILESOV (U.S.S.R.)

The Glaciologic group of the Department of Geography
of the Academy of Science of the Kazakh SSR.

SUMMARY

In this report the results of the preliminary analysis of the data on the temperature regime of the lower part of the Central Tuyuksu glacier in Zailisky Alatau (the North Tien Shan) are submitted. They are obtained by the action of a specially laid borehole, which was functioning from September 15, 1958 up to September 3, 1959. The period of the control made up a full year cycle. The temperature measuring was conducted three times a day by a platinum electrothermometre of resistance (3 T II-310a) set in the borehole at the depth of from 0,5 to 52,0 m., with the intervals of from 0,5 to 10 metres.

According to the temperature condition in the thick of the tongue-ice of the Central Tuyuksu glacier the following temperature zones are marked out (by G. A. Avsyuk) :

a) Ice tropozone — with the variable stratification of temperatures from the surface to the depth of 6 m. (with the crust of thawing - to 0,5 m.).

b) Stratozone of ice — with a constant stratification of the temperatures from 6 to 15 m.

c) Homothermic zone — with constant temperatures from 15 m and lower, up to the bed of the glacier.

The three temperature zones have clear run of the temperature during a year.

In this report we give middle temperature gradients, middle speeds of changing temperatures for the ice tropozone, the character of changing the sum of the inversional temperatures (in relation to their general stratification), which every 10 cm. of ice-thickness give.

The data of the distribution of temperatures in the bore-hole are co-operated with the motion speeds of the internal ice-strata.

The analysis of the temperature measuring in the bore-hole corroborates the supposition about a temperature regime of glaciers with a continental type of distribution of temperatures. (G. A. Avsyuk, Ye. N. Tsikin).

RÉSUMÉ

On trouve dans le rapport les résultats de l'analyse préliminaire des données sur le régime de température de la partie inférieure du glacier du Toujouksou dans l'Ala-Tau (Tien Chan septentrional) sur l'exemple des fouilles spéciales qui ont fonctionné du 15 septembre 1958 jusqu'au 3 septembre 1959. La période des observations comprit un cycle annuel complet. Les mesures de température furent faites 3 fois par 24 heures à l'aide d'électrothermomètres en platine (TII-310a) placés dans les fouilles à une profondeur de 0,5 à 52,0 m à des intervalles de 0,5 jusqu'à 10 mètres.

Dans l'épaisseur de la glace de la langue du glacier du Toujouksou central, on distingue d'après la température les zones de température suivantes (d'après G. Avsiouk) :

a) la tropozone de glace — avec une stratification alternative de la température de la surface jusqu'à 6 mètres de profondeur (avec une croûte de fusion d'une épaisseur pouvant atteindre 0,5 m);

b) la stratozone de glace — avec une stratification constante de la température de 6 à 15 m;

c) la zone homothermique — avec des températures constantes à partir de 15 m et plus bas, jusqu'au lit même du glacier.

Les trois zones de température ont une modification annuelle de température bien définie.

On trouve dans le rapport les gradients de température moyenne, la vitesse moyenne de changement de température pour les tropozones de glace, le caractère des modifications de la somme des températures d'inversion (par rapport à leur stratification générale pour chaque couche de 10 cm de glace).

Les données sur la distribution des températures dans les fouilles concordent avec les vitesses de progression des couches intérieures de la glace.

L'analyse des mesures de température dans les fouilles confirme l'hypothèse déjà émise concernant le régime de température des glaciers ayant une distribution de température du type continental (G. Avsiouk et E. Tzikine).

The study of the temperature regime of ice on the Tuyuksu glaciers, situated in the upper part of the river Malaya Almaatinka in the ridge of the Zailiysky Alatau had been carried out by the glacial expedition of the Department of Geography of the Academy of Science of the Kazakh SSR in 1957-1957 and is being continued until the present time. The largest valley glacier Zentralny Tuyuksu reaches 5,1 km and has in the cirque section a width of 1,3 km, on the tongue -0,5 -0,3 km. The highest point of the glacier basin is the peak Pogrebetsky (4218,9 m). The open end of the glacier's tongue lies at an altitude of 3370 m. The average inclination angle of the surface equals to 8-10°.

For the study of the temperature regime of intraglacial depths on different sections of the glacier of the Central Tuyuksu, in the summer of 1957, 4 pit-holes were bored 20 to 25 meters deep. In 1958, on the tongue of the glacier, another bore hole (No. 5) was bored that reached the bed moraine at the depth of 52,5 m. The characterization of the temperature regime of the glacier is given in this report according to the observations of temperatures in the bore hole No. 5. The latter was placed on the axis part of the glacier 600 m away from the extreme end of the tongue and 2300 cm from the foot of the back cirque wall at the altitude of 3480 m above sea level.

The highest average-monthly temperatures on the meteorological station Tuyuksu-2, that is situated directly near the bore hole are observed during four months of the year (from June to September) but they do not raise above +5,1°. The average monthly air temperatures for 1958 and 1959 on the station Tuyuksu -2 are given in table 1.

TABLE 1

Year	I	II	III	IV	V	VI	The year's average
1958	- 12.8	- 11.8	- 8.1	- 3.5	- 2.6	1.4	
1959	- 13.2	- 14.9	- 10.0	- 3.1	- 3.5	2.0	

Year	VII	VIII	IX	X	XI	XII	The year's average
1958	5.1	3.1	1.7	- 4.0	- 11.0	- 10.6	- 4.4
1959	4.1	4.7	4.6	- 2.5	- 10.6	- 13.3	- 4.6

The absolute maximum is noted here to be $+19.1^{\circ}$ (July 1959), minimum -28.2° (January 1959). The stable snow cover on the glacier is settled by the second half of September or the beginning of October. During the winter period the thickness of the snow cover varies from 100 to 160 cm with a density of 0.32-0.36. The tongue of the glacier is free from snow by the end of June, but the maximum melting takes place in July. During the ablation period in the region of bore hole No. 5, which is placed in the zone of maximum ablation, the glacier loses through melting as much as 150-200 cm of ice (converted into water).

The average duration of the melting period in the region of the bore hole lasts 70-80 days.

The observations of temperature on the tongue of the Zentralny Tuyuksu were carried out by means of platinum electrothermometers of resistance, frozen in the bore hole at depths of 0,5, 1,0, 2,0, 4,0, 10, 15,0, 20, 0,30, 0,40, 0,46,0, 48,0, 50,0, 51,0, and 52,0 m. During the observation period these depths diminished for account of ablation—in 1959 almost by 1,5 m. Besides that, in the same bore hole, at depths of 11,0, 17,0, 26,0, 34,0, 46,0 and 52,0 m were placed metal block-electrodes to obtain data on the motion of the inner layers of the glacier.

The measurement of the temperature were carried out, with the aid of a portable Wheatstone bridge (MTW) that was connected to the thermometer circuit at the moment of measuring. Accuracy of temperature measurements with platinum thermometers equals to $\pm 0,05^{\circ}$.

The readings off all thermometers were taken three times a day—at 7° , 13° and 19° hours. The bore hole functioned from 15,09, 1958 to 3,09, 1959). during this period over 11000 measurements of the ice temperature have been made.

As to the character of ice formed, the glacier Zentralny Tuyuksu, as well as all the Tien Shan glaciers, belongs to the infiltrate-congelation type, which is characteristic of continental districts. Therefore, the ice of the described glacier has a continental distribution of temperatures, that is, formed under the influence of rather considerable negative temperatures during the cold period of the year, and also a relatively low average temperature during the warmest month ($+5,1^{\circ}$) and negative yearly average temperature of $4-6^{\circ}$ below 0.

The analysis of the actual material of the temperature measurements on the tongue of the glacier Tuyuksu shows, that during the whole year, the entire 52 meter depth of ice has negative temperatures. Only in the upper layer of ice (at a depth of 0,7-1,0 m) which is intensively saturated by surface melt water during the warm season of the year, zero-temperatures are observed. In deeper horizons the following maximum temperatures have been noted: at a depth of 2 m $-0,6^{\circ}$, at the depth of 4 m $-2,0^{\circ}$ and at the depth of 6 m $-3,1^{\circ}$.

The results of temperature measurements throughout the entire depth of 52 meters are shown in the table 2.

From this table it follows that ice temperatures in the upper horizons down to a depth of 2 m and which are exposed to a direct influence of weather conditions, are very variable and follow the changes of the air temperature. The deeper horizons of the ice (below 2 m) have a rather smooth temperatures curve due to the air temperature having a lesser influence and with considerable delay. A strong cooling of the surface layer creates in deeper zones of the ice a dissimilar temperature field. This dissimilarity, connected with the penetration of the surface cooling of the body of the glacier, effects the value of the temperature gradient.

The calculations show that the absolute values of the temperature gradients diminish with depth, whereas in the upper part of the temperature field up to a depth of 2 m, sharper changes of the gradient correspond to a slight increase of depth, and inversely at greater depths (below 2 m), a slight decrease of the gradient corresponds to large changes of the depth. From table 2 it follows that on separate stages the strati-

TABLE 2
Average monthly ice temperatures by horizons

Period horizon m	15-IX	X	XI	XII	I	II	III	IV	V	VI	VII	VIII	Average
	30-IX												
0.5	-1.0	-2.4	-4.4	-5.4	-5.9	-7.1	-7.6	-6.5	-2.4	-0.5	-0.3	—	-4.2
1.0	-0.4	-1.9	-3.6	-4.6	-5.0	-6.1	-6.7	-6.1	-2.9	-1.5	-0.9	0.0	-3.8
2.0	—	-1.1	-2.7	-3.6	-4.3	-5.2	-5.7	-5.3	-3.3	-2.2	-1.5	-0.9	-3.4
4.0	-2.4	-2.4	-2.5	-2.7	-3.0	-3.6	-4.6	-4.4	-4.0	-3.5	-2.6	-2.1	-3.2
6.0	-3.2	-3.2	-3.2	-3.2	-3.2	-3.2	-3.3	-3.4	-3.4	-3.4	-3.2	-3.1	-3.2
10.0	-2.4	-2.4	-2.4	-2.4	-2.5	-2.5	-2.6	-2.6	-2.6	-2.6	-2.5	-2.5	-2.5
15.0	-1.7	-1.7	-1.7	-1.7	-1.8	-1.8	-1.8	-1.8	-1.8	-1.8	-1.8	-1.8	-1.8
20.0	—	-1.3	-1.3	-1.3	-1.3	-1.4	-1.3	-1.4	-1.4	-1.4	-1.4	-1.4	-1.4
30.0	—	-1.0	-1.0	-0.95	-1.0	-0.95	-0.9	-1.0	-0.95	-0.95	-0.95	-0.95	-0.95
40.0	—	-0.9	-0.95	-0.9	-0.9	-0.9	-0.9	-0.9	-0.9	-0.9	-0.9	-0.9	-0.9
46.0	—	-0.2	-0.2	-0.2	-0.2	-0.2	-0.4	-0.4	-0.5	-0.6	—	—	—
50.0	—	-0.1	-0.1	-0.1	-0.1	-0.2	-0.2	-0.2	-0.3	-0.4	-0.5	—	—
52.0	—	-0.1	-0.1	-0.1	-0.1	-0.1	-0.2	-0.2	-0.3	-0.4	-0.5	-0.7	—

fication of the raise of temperatures by increase of depth (a direct or "winter" stratification) is replaced by an opposite one ("summer" stratification). A similar change of stratifications takes place in the upper part of the temperature field at a depth of 6 m by the approach of the winter cold and at the beginning of the warm period of the year. To have a clearer notion about the changes of the temperature stratification by the depth of all horizons, where a change of stratification takes place, we summed up the invernal temperatures that cover an ice layer by a thickness of 1 decimeter (table 3).

TABLE 3

Summaries of the average daily invernal temperatures by all upper ice horizons.

Horizon (m)	Sum of invernal temperatures, that cover an ice layer of 1-dm	
	From 20.IX.58 to 31.I.59	From 1.V.59 to 31.VIII.59
1.0	0°	12.4°
2.0	0.6°	6.5°
4.0	1.8°	6.4°
6.0	3.6°	2.4°

As this table shows, the upper ice layers are strongly warmed in the summer, and the maximum sums of invernal temperatures fall on the warm period (the temperature gradient changes at that time changes signs). Deeper than 6 meters and all the way down to the bottom moraine a direct stratification of temperature increase takes place, by increase of the depth, during the whole year.

As it has already been pointed out, the upper ice horizons (by a depth up to 2 m) are distinguished by an extremely unstable temperature regime, accompanied by rather frequent temperature changes (towards raise or fall). In the deeper layers of the glacier the amplitudes of the temperature variations are being noticeably decreased. The deepest ice horizons are being characterized by slight (at the depths of 6-10 m) and even zero (deeper than 10 m) temperature variations during the entire year's observation cycle (table 4.)

With the increase of the depth decreases also the rate of temperature changes, reaching at a depth of 10 m 0,02° within 24 hours, but at the depth of 15 m -0°.

Usually the study of the temperature condition of glaciers is accompanied by defining temperature zones, that are notable for specific peculiarities of the course of the temperatures.

The division of the depth of the glacier into zones is generally made according to character of the temperature stratification. Proceeding from above quoted material, the entire depth of the glacier tongue may be divided, according to its temperature regime, into three principal zones (G. Avsyuk, 1954, 1955, 1956; E. Tsykin, 1957): the surface, the middle, and deepest.

1. The superficial, or tropozone of ice spreads from the surface to a depth of 5 m; 2. The middle, or stratozone of a 9 meter-thickness is placed at the depth from 5 to 15 m; 3. The deepest zone is considered below 15 m all the way down to the bed-moraine.

TABLE 4

Regime of temperature variation in the upper horizons of the glacier

Horizon (m)	Number of changes of temperature varia- tion symbol	Average duration of of changes of one symbol (days)	Average changing rate of temperature (in degrees within 24 hours)
0.5	88 to 313 days	3.5	0.23
1.0	45 in 320 days	7.1	0.15
2.0	17 in 227 days	13.3	0.09
4.0	7 in 332 days	47.7	0.06
6.	1 in 263 days	263	0.04
10.0	0 in 278 days	278	0.02

The superficial zone, or tropozone, is characterized by the regime of variable temperature stratification (Fig. 1). This regime is formed by the direct influence of air temperature fluctuation upon the glacier, and by warming or cooling of the upper ice layers, and also by the thermo-influence of the surface melt water that penetrates into these horizons. The influence of these processes, connected with the motion of the lower depths of ice, in comparison with the above said factors, effect very little the temperature regime of the tropozone.

The temperature stratification within the tropozone varies with the seasons of the year as follows. During the warm season of the year, owing to high air temperature, intensive solar radiation and melting, the upper ice layers are strongly warmed, and therefore with an increase in depth the air temperature sinks, i.e. a summer stratification is being created. According to the extent of warming of the upper ice horizons the minimum temperature sinks down to the lower layers of the tropozone and in the middle of the warm period (July-August) settles down on the lower edge of zone i.e. at the depth of 6 m. In the winter months, in view of the spreading of the cold, the high ice temperatures of the surface horizon sink rapidly, and the minimum temperatures are again transferred, but this time from the lower layers to the upper ones, according to the cooling extent of the latter. The ice temperatures, inherited from the warm period of the year, at depths of 4-6 m are preserved for a rather long time.

In February, March and April, throughout the entire depth of the ice tropozone there takes place a regular winter temperature stratification, during which the minimum temperatures are on the ice surface. In the spring and autumn, in the surface zone there is observed a mixed temperature stratification. Within the limits of the tropozone, according to the character of temperature changes by depth and average temperature gradient, it is possible to single out two temperature subzones: 1. a subzone of variable stratification, conditioned by intraseason weather fluctuations and 2. a subzone of variable stratification, conditioned by change of the seasons of the year.

The upper subzone is bedded at a depth of up to 2 m from the surface. Above 2 m the ice temperature is subject to rapid and sharp fluctuations. It is here that the maximum amplitudes of the temperature fluctuation had been fixed. In particular

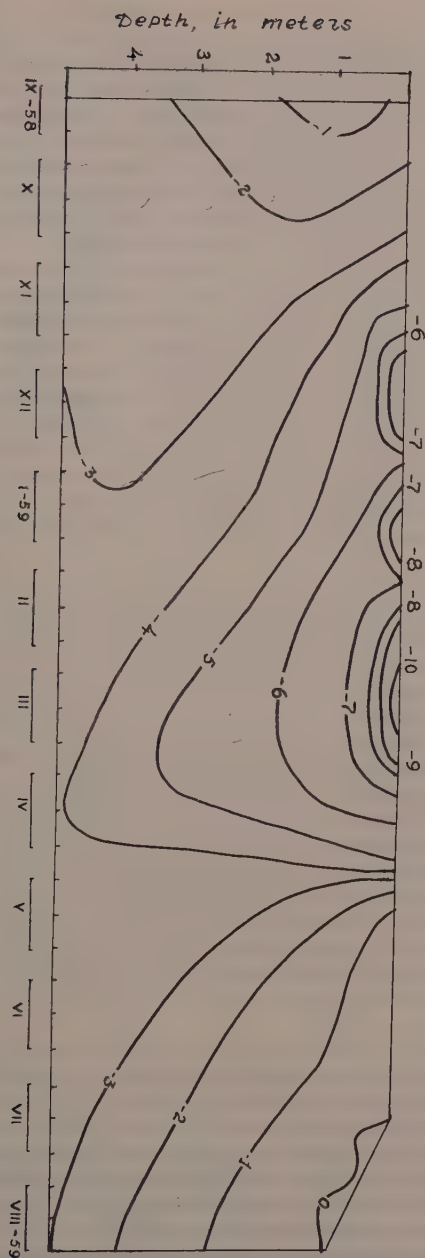


Fig. 1 — The course of the average ice temperature by decades, up to a depth of 5 m on the tongue of the glacier Zentralny Tuyuksu (from 20.IX.1958 to 31.VIII.1959).

the yearly amplitude reaches $8-10^{\circ}$. The ice temperature of the subzone is strongly dependent on the depth, and the temperature gradients have the most significance (averaging during the observation period $0.004-0.008$ of a degree on each centimeter). The maximum gradient of the change of temperatures was noted in the winter (January-March), when it reached 0.02° on each centimeter and more. It is the winter when the largest diversity of temperature is noted within the limits of the subzone—up to $5-6^{\circ}$. The ice subzone considered here is characterized also by a maximum speed of the temperature change ($0.09-0.023^{\circ}$ during 24 hours. The influence of the daily temperature fluctuations of the air layer adjacent to the glacier effects a depth of $0.7-1.0$ m, i.e. within the limits of the so called "thawing crust".

The subzone of variable temperature stratification, conditioned by the change of the seasons of the year, has a thickness of 4 m and is bedded on depths from 2 to 6 m. In the temperature regime of this subzone are distinctly shown the season changes of the temperature conditions of the surrounding air and solar radiation. Below 2 m are observed to be rather slow in comparison with the upper subzone changes of ice temperature. The speed of these changes is not quite dependent on the depth and varies within 0.04 to 0.09° within 24 hours. The significance of the temperature gradient noticeably decreases as the depth increases. The annual amplitude of air fluctuations within the limits of the subzone equals to $1.5-2.5^{\circ}$, but near the lower edge of the subzone it decreases to $0.4-0.5^{\circ}$. The maximum temperature diversities between separate horizons of the subzone do not exceed $2.0-2.5^{\circ}$ even in the winter period.

The middle zone or ice stratozone is characterized by a regime with a constant temperature stratification. The temperature regime of the stratozone is formed under the influence of both the outer factors, the influence of which predominates in the upper layers of the zone, as well as the intra-glacier processes (ice motion, etc.) and the influence of which predominates in the lower part of the zone.

Within the limit of the stratozone the ice temperature is, during different seasons of the year, always negative and varies depth from -3.2° to $-1.7-1.8^{\circ}$. The influence of the season temperature fluctuations of the air layer adjacent to the glacier upon the temperature condition of the stratozone ice, shows very faintly and with a considerable delay (3-4 months). By increases in depth the "waves" of cold and warmth show to a lesser extent and approaching the lower edge of the zone, die down almost completely. The temperature curve at depth in the zone is an extremely smooth one. The general character of the direction of the thermoisohypsos is to be drawing nearer to the parallel with the surface of the glacier, which proved by the distribution of temperatures within the limit of the stratozone in different seasons of the year (Fig. 2). The temperature diversities of different layers of the zone are not large—within $1.4-1.6^{\circ}$, the gradients of temperature changes are also not large and amount to $0.0014-0.0020^{\circ}$ on each centimeter.

The zone of depth is characterized by the temperature regime that is formed under the influence of the energy from intraglacial processes and also under the influence of the temperatures, that were inherited by the ice at its formation out of the solid precipitations in the accumulation region of glacier. The ice temperatures within the zone are distinguished by the constancy of their significances. The isothermal character of the temperatures particularly distinctly traced in the upper and middle part of the zone, at the depths of 15-40 m, where during the yearly observation period the temperatures did not change more than by 0.1° .

However, on the whole, the distribution of the temperatures in zone of depth is distinguished by some other regularities, that could hardly have been expected. First of all, in the ice of the zone is marked a sufficiently expressed tendency to an increase of temperature by the depth (from -1.7° to -0.7°), although the average temperature gradients are decreasing to $0.001-0.0001^{\circ}/\text{cm}$. Particular consideration

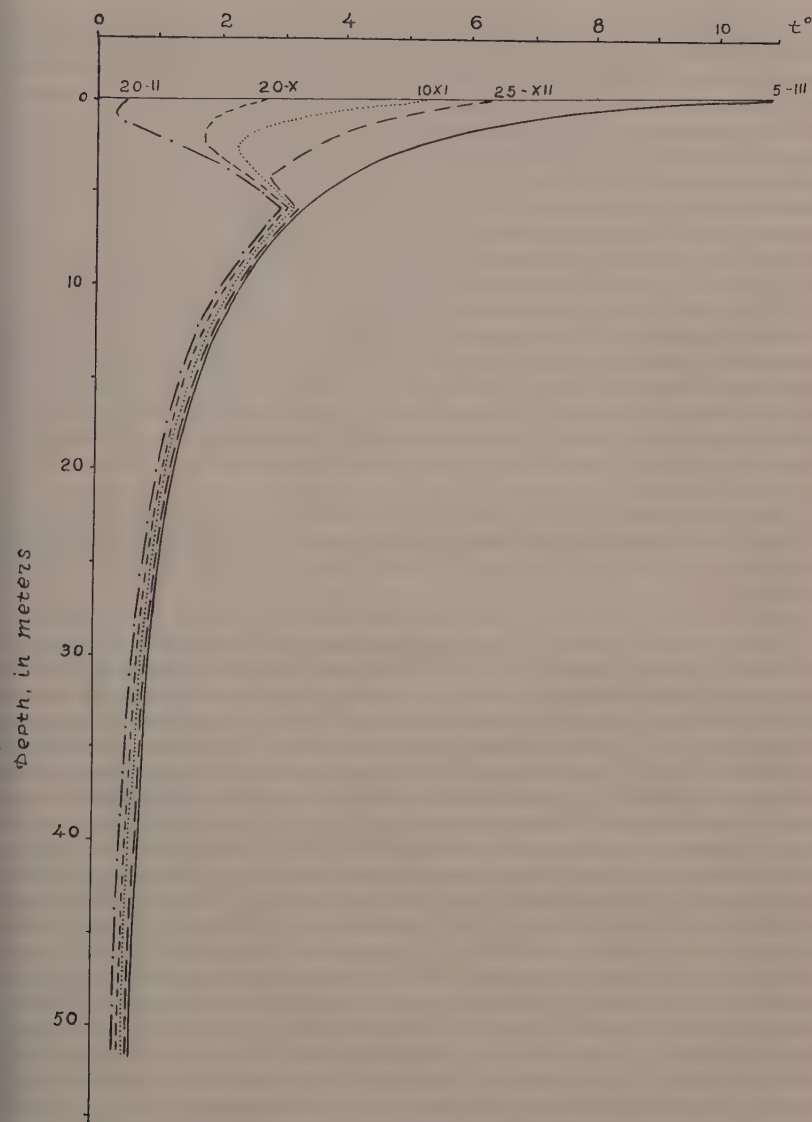


Fig. 2 — The distribution of the temperatures with depth in the ice on the tongue of the glacier Zentralny Tuyuksu (1958-1959).

is deserved of the distribution of temperatures in the lower part of the zone at the depths of 46-52 m. Here attention is attracted to the fact that the temperatures from October 1958 to February 1959 are equal, throughout the entire 6 m thick ice adjacent to the bed moraine, to about -0.1° . At the beginning of March there begins a gradual fall of the temperature that lasts until the end of the observation period (the end of August 1959), when it reached -0.6 — -0.7° . Such sharp change of temperature for the depth — one may be explained by the following factors.

On August 22 1958, on completion of boring of the pit hole, the latter (after lowering a cable with the thermometers), was flooded with water that for a certain space of time warmed up the ice adjacent to the pithole. It required about 30 days until the water in the upper half of the borehole froze and adapted the temperature of the surrounding ice. But in the lower part of the borehole (below 46 m) the water, apparently did not freeze till the end of October, when, at last, the thermometers showed a negative temperature -0.1° . This temperature was maintained on one and the same level during the four winter months because during this period there was a maximum motion of the glacier tongue. Owing to the slowing-down of the speed after February (and also due to a certain extent by the time factor), the ice formed in the bore hole by freezing of the water begins to adapt, a great deal faster, the temperature of the surrounding ice layers. Apparently the temperature -0.6° — -0.7° , that has been fixed at the depths of 46-52 m already at the end of the observation period, is the very temperature, which is natural to this zone at the given depths. Unfortunately, at the beginning of September 1959, all the thermometers went out of use (owing to a rupture of the feeding wires, caused by irregularity of the distribution of motion speeds at separate depth of ice layers in comparison with the surface speed of the glacier). Therefore, there was no possibility to trace the further source of the ice temperature in the bore hole.

It is known that in glaciers with infiltrate-congelation (continental) temperature regime, the ice motion, which has a plastic character, the dynamic intra-glacial processes cause a noticeable change of temperatures, especially in the deep layers of glaciers.

As G. Avsyuk (1956) shows the distribution of quantity of heat that is formed by the motion of ice on a vertical line, is proportional to the diversities of speeds of differential motion of separate ice layers, and therefore, the maximum quantity of heat ought to enter into the lower bed ice horizons, where it is possible to rise to near-bed melting temperatures. By the raising from near the bed of the glacier to its surface, the quantity of heat, and this means also the warming of the glacier, will diminish. At depths above 15 m the influence of the motion processes is not so large as more considerable influence is rendered here by the season and yearly temperature fluctuations of the external air. However, in connection with the small motion speeds of the glacier Zentralny Tuyuksu in the considered part of the tongue (5.8 m in a year at the surface the raise of the ice temperature, conditioned by the motion, is not strictly proportional to the diversities of the differential motion speeds, but has a more smoothed and regular character. The regularity of rise of the ice temperatures by increases of the depth is explained by the thermoexchange between separate layers, which, owing to the low flow speeds, has time to smooth out considerably the temperatures in the entire depth of the ice. On the tongue of the glacier Zentralny Tuyuksu where the motion speed at a depth of 52 m is reduced by 34% in comparison with the surface speed (B. Borovinsky, K. Makarevich, 1959). This smoothing out of temperatures leads to such a condition that the ice temperatures rise almost equally in the depth-zone of the glacier. The above said is well illustrated by Fig. 3. On the curve of the relative transference of points (A), attention is attracted to the irregularity of the distribution of change of motion speeds of separate ice horizons in relation to the surface speed. The irregularity of change of speeds between separate ice horizons is particularly clearly shown by the speed gradients, that were calculated by five meter-distance sections (curve B). The position of the section "a¹" of the temperature gradient (curve C) shows that the change of the temperatures within the tropozone (from the surface down to a depth of 6 m), is dependent on seasonal and inter-seasonal weather fluctuations, occurs extremely sharply and is almost not dependent on the motion speed. The sections "b" (of the speed gradients curve B) and "b¹" (of the temperature gradient's curve C), are, by their position, quite similar

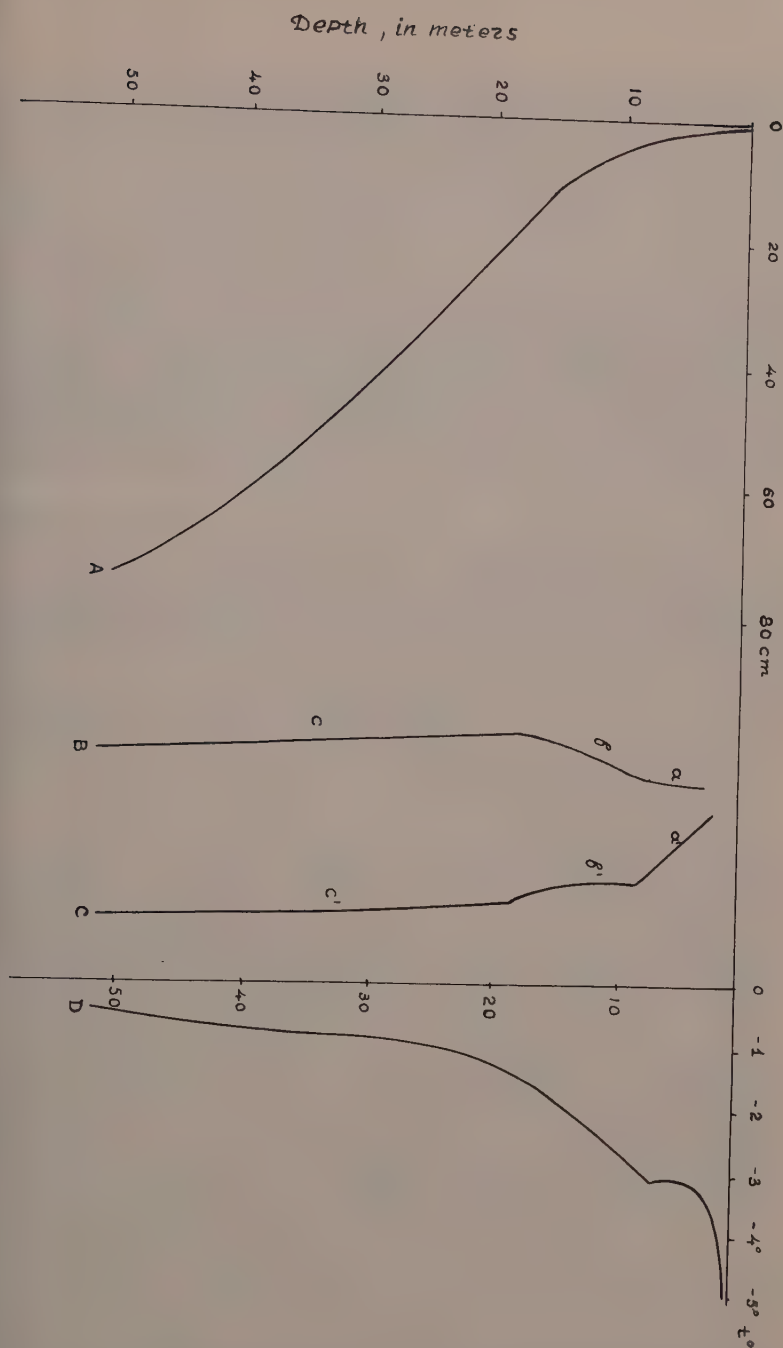


Fig. 3 — Change of motion speed and of temperatures with depth in the ice of the glacier Zentralny Tuyuksu according to data of a 52-meter deep borehole (from 22.VIII.1958 to 27.2.1959).

- A — delay, in cm;
- B — gradient values of motion speed;
- C — the average ice temperature during the same time;
- D — values of temperature gradients.

to each other, i.e. on the change of the ice temperature at the depths from 6 to 15 m, to a certain degree also begins to tell on its motion. Lastly the position of the section "c" and "c¹" of both curves certify that the ice temperatures at a depth of 15 m and lower, change (increasingly) almost entirely according to the changes of the motion speed. The quantity of warmth that is formed by the motion on different ice horizons has not yet been calculated, but it may now already be established with confidence that the entire glacier, in relation to the distribution of motion speeds, can also be divided into three principal zones—the surface, middle and deep zone, the thickness depository limits, of which correspond to thickness and depository limits of the temperature zones.

The results of measurements of temperature and motion speed of the lower depths of the ice on the tongue of the glacier Zentralny Tuyuksy in 1958-59 lead to the following conclusions:

1. Owing to the fact that the surface of the lower part of the glacier is composed of ice so dense that is almost completely impervious to melt water, the ice temperature at every depth in all seasons of the year has a negative value, with the exception of the "melting crust".

2. The ice on the tongue is subdivided into 3 principal temperature zones: the surface, the middle, the deep zone and each differ from each other by the temperature regime and by the influence of the factors that form this regime.

3. A determinative influence upon the temperature regime of the depth-layers is rendered by the dynamic intraglacial processes; first of all is the motion of glacier which is also divided into 3 principal zones.

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ON THE MEASUREMENT OF GLACIER FLUCTUATIONS

Richard FINSTERWALDER, Munich

SUMMARY

For studying the change of climate it seems necessary to measure glacier fluctuations in a distinct way. Chiefly measurements which relate to the change of height of glacier surfaces and of the snow-line are of interest and importance. Such measurements have been executed and evaluated since some decades in the Eastern Alps. Especially more exact measurements in the last decade 1950-59 are described. The results are combined with those of former decades. — For getting a better insight into climatic changes of shorter and longer period one has to measure glaciers of different size and glaciers which react quickly or in longer times. — In order to obtain useful material for the World data centers it is suggested to observe a certain program for an international cooperation.

RÉSUMÉ

Pour étudier les variations du climat il est nécessaire de mesurer les fluctuations des glaciers en employant certaines méthodes. Des mesures de la variation de la hauteur des surfaces des glaciers et de l'altitude de la ligne de névé sont surtout intéressantes et importantes. De telles mesures ont été exécutées depuis plusieurs décades dans les Alpes orientales. En particulier des mesures plus exactes de la dernière décade 1950-59 ont été décrites. Les résultats sont combinés avec les résultats des décades antérieures. — Pour atteindre une meilleure connaissance des variations du climat plus courtes et plus longues il est utile de mesurer des glaciers d'étendues différentes ainsi que des glaciers qui réagissent soit rapidement soit lentement aux influences du climat. — Pour atteindre des résultats utiles pour les World data Centres on propose de suivre un certain programme de collaboration internationale.

The great number of glaciers and their variety in form of appearance make the investigation into glacier fluctuations a difficult task. Such investigations will be really teeming if they can be connected with the influence of climate and weather.

For this purpose mainly figures and observations are important which relate to the fluctuations in levels of glaciers:

1) Fluctuations of the surface levels over the whole glacier and in separate high-zones.

2) Fluctuations in the height of snow-line on the glaciers.

3) With certain reservations also fluctuations in the levels of the glacier tongues.

These changes in height can be connected in different ways with meteorological elements and the run of climate: Fluctuations in surface levels naturally first of all with precipitation and ablation, the latter depending in complicated manner from meteorological elements. At a preliminary interpretation of the change of snow-line mainly the vertical gradient of airtemperature could be taken into consideration. But it has to be emphasised, that it is the task of meteorologists and climatists to interpret observed glacier fluctuations with the help of their sciences.

Except from the mentioned fluctuations in level, glacier fluctuations can be described to some extent by the change of areas covered by glaciers, the retreat and advance of the very tongue-ends in directions of the respective glacier bed. These characteristics, especially the last one, can be surveyed relatively easiest. The variations of tongue-ends are—as it is well known—systematically observed by the Swiss Commission on Glaciers in Switzerland ⁽¹⁾ and in Austria ⁽²⁾ by the Austrian Alpenverein since many years in a deserving manner. However direct numerical conclusions on meteorological elements or on climate can not be gained in this way.

The relations between glaciers fluctuations and change in climate could be evaluated best by meteorological and geodetic measurements and interpretation of the total snow and ice economy including the vertical and horizontal movements of the glacier. In this respect pioneer work has been done by W. von Ahlmann⁽³⁾; the most intensive examination of one glacier was executed by O. Schimpp at the Hintereisferner⁽⁴⁾. But such thorough examinations are so timeconsuming and difficult that they hardly can be executed at several glaciers for an extended period of time.

1. INVESTIGATIONS IN THE EASTERN ALPS 1856-1950

The undersigned therefore has tried to evaluate numerically—based upon photogrammetrical surveys of a number of glaciers in the Eastern Alps—mainly, the change in levels of glacier surfaces and of the snow-line in the time between 1920 and 1950. The applied methods have been described in detail in⁽⁵⁾, shortly also in⁽⁶⁾.

Regarding the change in level of glacier-surfaces also the time since the last climax in glacier advance at about 1856 could be incorporated. The result is a negative change in levels, i.e. a medium loss of height of all glacier surfaces. The average of this is for the periods:

Period	Number of Years	Ablation/Year	Ablation/Period
1856-1890	34	0,60 m	20,4 m
1890-1920	30	0,30 m	9,0 m
1920-1950	30	0,61 m	18,3 m
			<hr/> 47,7 m

Regarding the change in level of the snow-line, at eight different glaciers for the period from 1920-1950 a mean of 62,5 m rising for this line was found. When considering the fact, that at big glaciers the tongues are following more slowly the glacier retreat, a higher figure of about 90 m was found. This figure has been directly calculated at three typical glaciers: at Sulzenau-, Waxegg- and Schlegeisglacier. The loss in surface per year was 0,56% of the area of glacier.

2. RECENT INVESTIGATIONS FROM 1950-1959

Whereas the above mentioned investigations refer to periods of about 30 years and the used observations had also an interval of mostly 30 years each, it was now to be tried, to take measurements in shorter intervals in the period from 1950 to 1960. The shortest possible time interval for such geodetic-photogrammetric surveys is one year, and the observations are done at the time of the snow minimum in the latter summer. This interval of one year during the whole period could only be fulfilled at two glaciers; the Waxegg-glacier in the Zillertal and the «Schneeferner» in the Wetterstein-mountains. Mainly the Waxegg-glacier is interesting and shall therefore be discussed with preference. The following table shows all measurements executed. These measurements comprehend either the total glacier or only the tongue.

TABLE 1

Analysed glacier (Eastern Alps).

Site of glacier	Glacier	Surface km ²	
Zillertaler-Gruppe	Waxeegg	3,8	1950-1959 all years
	Schlegeis	5,5	1950-1954, 1959
	Horn	3,8	1950, 1954, 1958
	Schwarzenstein	4,6	1950, 1954, 1958
Stubai	Grünau	1,8	1950, 1959
	Sulzenau		
	Fernerstube	4,4	1950, 1959
Ötztal	Gepatsch	17,5	1953, 1956, 1957, 1958
	Hintereis	10,0	1953, 1956, 1959
	Knötteln	0,5	1959
	Hochvernaglwand	0,8	1959
Wetterstein	Schneeferner	0,36	1951-1959 all years

As to be seen from the table, the examined glaciers have an average size of a few km², only the Gepatschferner with 17 km² and the Hintereisferner with 10 km² exceed this average. Also it can be seen from the table, that the large glaciers have not been surveyed each year, since that would have caused too much expenditure and work. The smallest glacier is—with only 0,36 km² in surface—the Schneeferner; this has been included since it is the northernmost glacier of the Alps.

The mentioned glaciers of only few km² in area are best suited for investigations into present glacier fluctuations and thus for climate fluctuations of short periods. With 2-3 km in length they through differences in level of 1000 to 1400 m. The ice at the tongue-ends, coming from the «Bergschrund» and having passed the glacier at the bottom, is approximately 60 to 100 years old. Since the ice apart from the bottom is essentially younger and since glacier-waves caused by fluctuations of snowfeeding, are running through the glacier much quicker than with the average speed of 30 m/year, it can be assumed that these glaciers adjust themselves quickly to changed climate conditions, that means within a few years, partly even immediately. For this adjustment the following scheme can be given:

1) Height of snow-line, level of tongue-ends and area of glacier-surface are following changes in climate within a very few years.

2) The mean change in height dh_m of the whole glacier can be assumed to react directly within a year on the total influences of weather.

For glaciers still smaller these statements are even more valid. But small glaciers have the disadvantage that in cool and wet summers the winter snow is not melting off sufficiently, so that then they cannot be surveyed.

Naturally the big glaciers—Gepatsch and Hintereisferner—have the tendency of following but slowly in their conduct fluctuations in climate. Especially the tongue-ends are lying in a too low altitude at glacier retreats. This fact concerns at present mainly the Hintereisferner. Its tongue is a remainder of the last climax in glacier advance of 1856, it is lying in a nearly horizontal valley and is fed in its lower part insufficiently, as has been proved by Schimpp (4). This tongue is moving little and melting with especially high figures of ablation. Numerical results of changes in height

derived from this glacier therefore cannot be generalized; they have been put in brackets in the following tables.

3. WAXEGG—GLACIER 1950-1959

The Waxegg-glacier is a typical glacier of only a few km² size. Fig. 1 shows its position of 1959, which at the tongue has retreated considerably in comparison with 1950. On the photo also the side moraine of the climax 1856 is visible; also

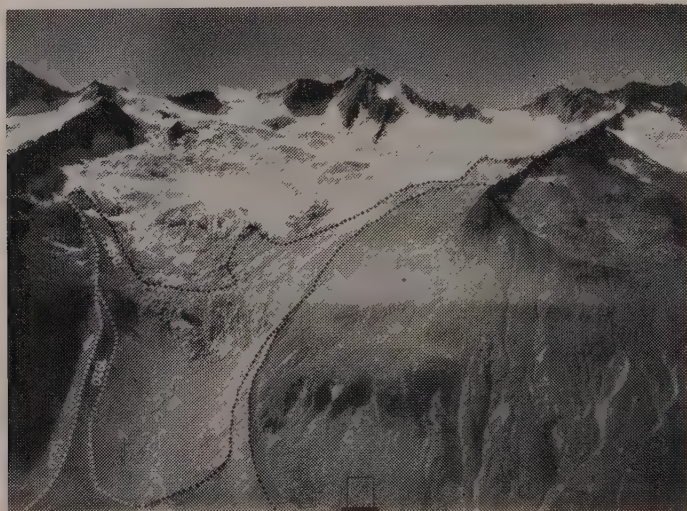


Fig. 1 — Waxegg-glacier 1959 (photo).

TABLE 2

Changes in level for each year and for zones of 100 m height difference.

height zones	areas (in ha) of height zones	Waxegg Glacier 1950...1959									
		1950	1951	1952	1953	1954	1955	1956	1957	1958	1959
3 200	23,0	m	+ 3,4	- 2,0	- 1,1	0,0	+ 1,5	(+ 1,0)	(+ 1,0)	- 0,5	- 2,0
3 100	33,0	"	+ 4,1	- 1,3	- 1,2	- 0,7	+ 1,5	(+ 0,6)	(+ 0,5)	- 0,1	- 1,7
3 000	38,4	"	+ 3,9	- 0,9	- 0,2	+ 0,2	+ 2,0	(+ 0,7)	(+ 0,5)	- 0,9	- 1,0
2 900	52,3	"	+ 2,4	0,0	- 0,1	- 0,2	+ 2,4	(+ 0,8)	(+ 0,8)	- 0,9	- 0,3
2 800	51,6	"	+ 1,2	+ 0,9	+ 0,2	- 1,7	+ 2,8	(+ 1,2)	(+ 1,1)	- 0,9	+ 0,6
2 700	44,5	"	- 0,6	+ 0,6	- 0,1	- 1,1	+ 2,2	(+ 1,1)	(+ 1,2)	- 0,3	+ 0,8
2 600	30,4	"	- 1,1	- 0,3	- 0,5	- 1,0	+ 0,3	(+ 1,3)	(+ 1,5)	+ 0,1	+ 0,8
2 500	12,4	"	- 2,1	- 2,9	- 1,4	- 1,1	- 1,1	(+ 1,0)	(+ 1,1)	- 0,5	+ 0,5
2 400	4,6	"	- 2,8	- 5,6	- 6,6	- 1,8	- 2,3	(- 1,2)	(- 1,1)	- 0,6	"
2 300	1,2	"	- 2,8	- 8,7	- 1,5	- 1,6	- 1,9	(- 1,1)	(- 1,1)	(- 0,6)	"
2 200	mean dh _m		+ 1,4	- 0,4	- 0,4	- 0,7	+ 1,8	(+ 0,9)	(+ 0,9)	- 0,6	- 0,2

the position of 1920 is to be recognized from the bright and nearly bare moraine. Fig. 2 shows the photogrammetric contour plan with glacier boundary lines and contours of 1950 and 1959. These contour plans were plotted with the highest possible accuracy. Base for the plotting were stereo-photos, taken with the light field equipment TAF of ZEISS, picture-size 13/18 cm. This phototheodolite TAF with its good preci-

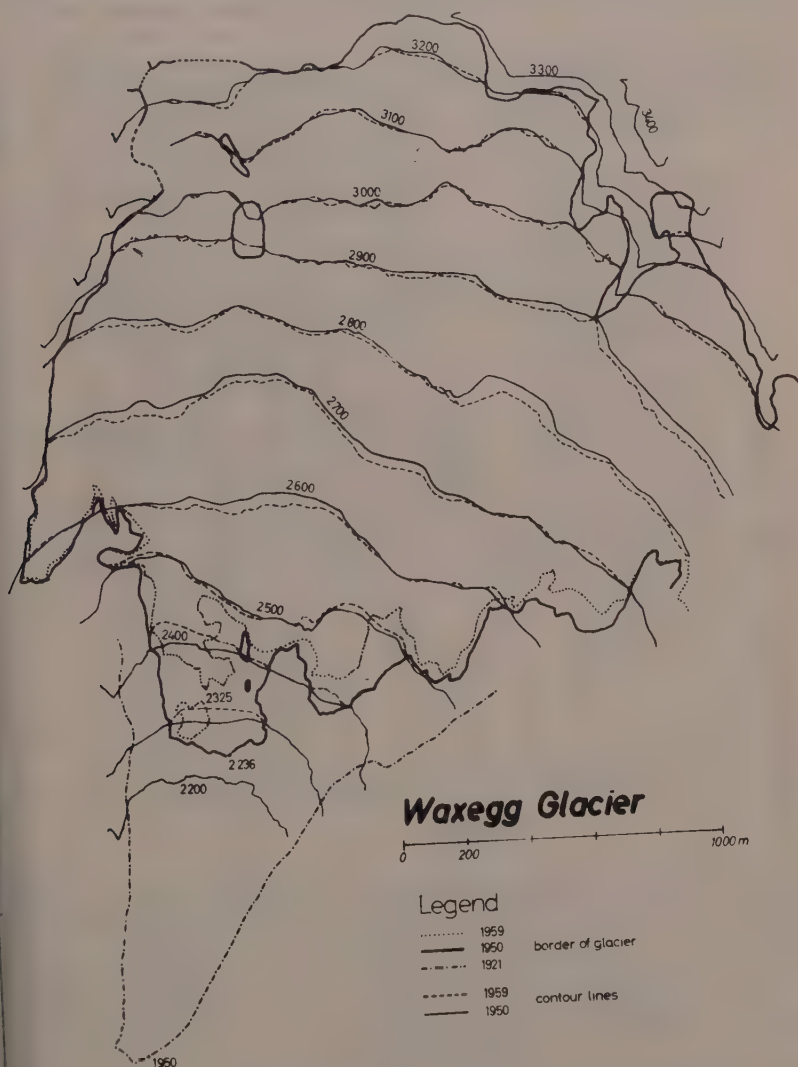


Fig. 2 — Waxegg-glacier contour-plan.

sion at only 10 kg weight (including tripod), was of great value for us. The photos taken each year were plotted on a transparent stable foil. Great accuracy and care was applied, since the annual change in glacier surface is small; consequently the contours are shifted only by small amounts from year to year. From the shift of con-

tours the changes in level of each zone of height and the mean changes in level dh_m for the whole glacier were calculated; the derived values for all zones and each year are given in table 2.

The fluctuations of the glacier surface in the different years and heights is shown by the diagram fig. 3, which has been drawn similar to table 2, but contains also the summed up changes in level since 1950. Lines of equal change in level have been drawn.

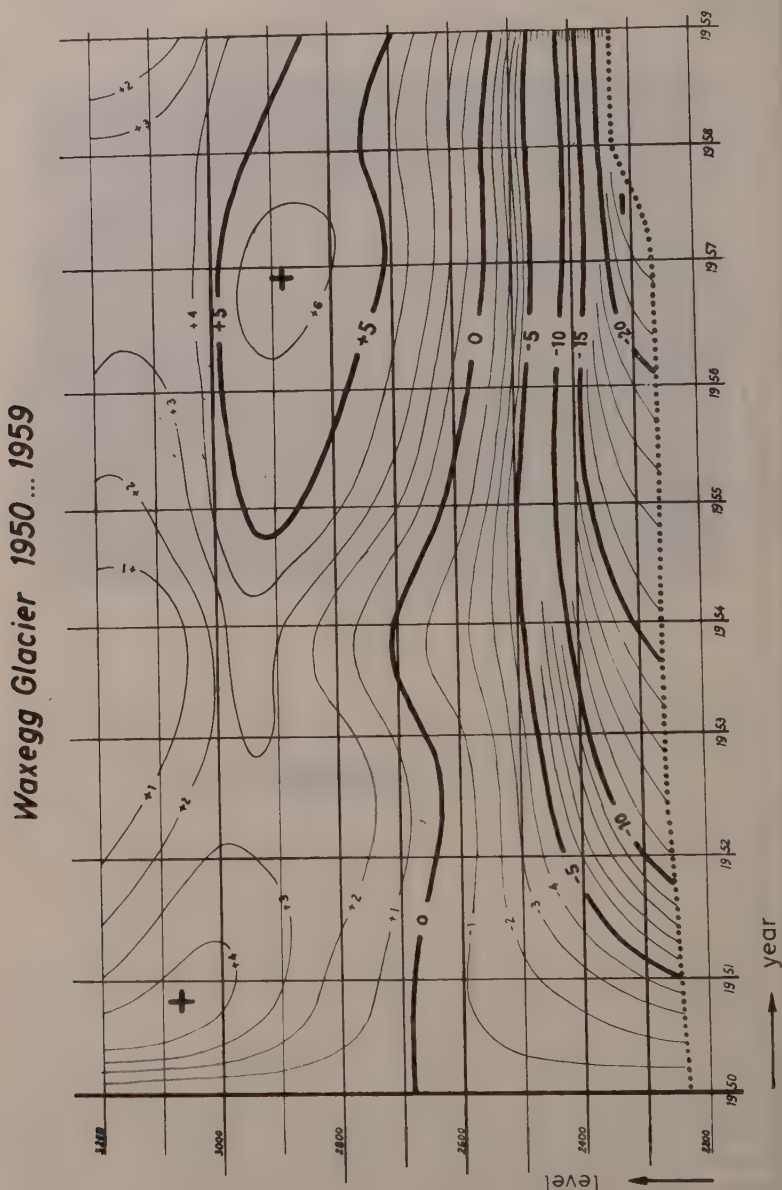


Fig. 3 — Lines of equal changes in glacier surface levels since 1950; measuring unit: meter.

It is to be seen, that since the year 1954-55 in an altitude of 2900 m an accumulation of more than 6 meters has taken place. By this a glacier wave is formed with the tendency of flowing down; this can also be judged from the inclination of the zero-line and the isolines. The tongue is melting off strongly. There the amounts of ablation are exceeding 20 m. The tongue end has risen since the last 9 years by a total of 100 m.

The comparison of the changes in height at Waxegg-glacier with the glacier retreat from 1920 to 1950 is very instructive. The values for the changes in height dh_m of the total glacier, the changes in level of the tongue end and the level of the snow line have been entered into fig. 4. Most important are the values dh_m for the change in height of the whole glacier. It is to be seen, that since 1951, the surface in total has risen although the tongue is still strongly melting off. Probably this is a consequence

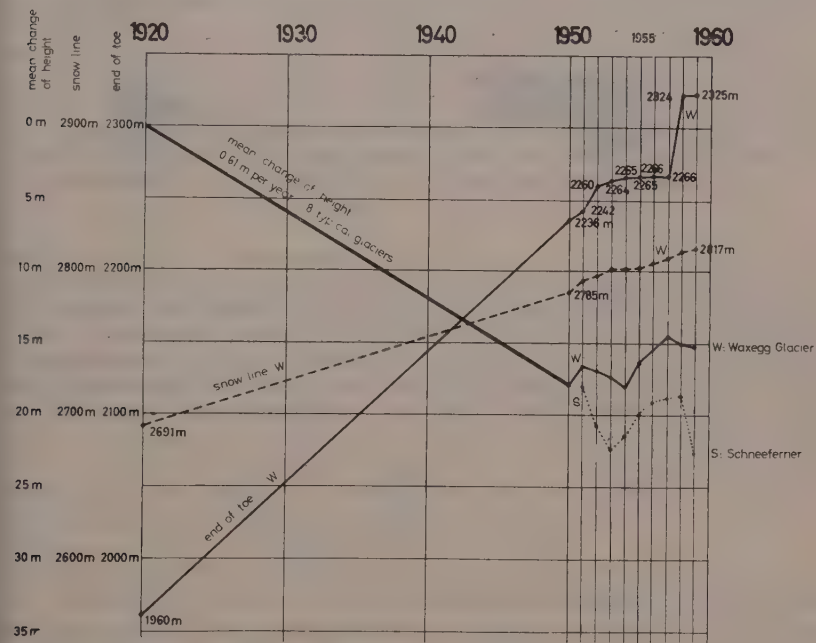


Fig. 4 — Diagrams.

of the then cool and wet summers, in which also radiation was low. The fine summers of 1958 and 1959 result at Waxegg-glacier in a slight decline of height over the total surface area. Similar is the progress of the dh_m values at the Schneeferner; this glacier has been influenced in 1959 as the most northern glacier of the Alps by the great hot-period being effective in the areas north of the Alps. In this summer the glacier lost 4 meters in height.

In table 3 a summary on the changes in height dh_m of all glaciers observed from 1950-1959 is given... From this table it can be seen:

- 1) Within single mountain areas glaciers behaved substantially in an equal manner.
- 2) The extent of glacier retreat between 1950-1959 slowed down at least in comparison with the period 1920-1950.

This can be derived from the fact that the average loss of glacier surfaces was reduced from 0,61 m/year to 0,13 m/year.

TABLE 3

Mean annual changes dh_m in level of some glaciers (in meters)

Glacier	1950	51	52	53	54	55	56	57	58	59	Mean
ZILLERTALER											
Waxegg	+1.5	-0.4	-0.4	-0.7	+1.8		+0.9		-0.6	-0.2	+0.31
Schlegeis	+0.41				+0.35						+0.36
STUBAIER											
Grünau						-0.24					0.24
Sulzenau						-0.20					-0.20
OTZTALER											
Hintereis						(-1.06)					-1.06
Gepatsch						-0.15		+0.30	+0.34		+0.04
Schneeferner		-28	-16	+0.9	+16	+0.9	+0.2	+0.1	-4.0		-0.59
	1950	51	52	53	54	55	56	57	58	1959	-0.13

In detail one can see from table 3 the weather-influence on the glaciers for all measured years and periods since 1950. It is interesting that the glacier retreat from 1920-1950 is still effective at the tongues and at the snow-lines. All glacier tongues are rising, i.e. retreating. Also the calculated snow-line is rising chiefly in consequence on the loss of area at the tongues. The values given in Fig. 3 for the rise of snow-line and tongue-end for Waxegg-glacier can be found in a corresponding amount also at other glaciers of the same type. Some values will be shown below.

TABLE 4

Rise of snow-line from 1950-1959

	Height of snow-line			Rise
	1950	1958	1959	1950...1959
Waxegg	2785 m	—	2821 m	36 m
Schlegeis	2800	—	2824	24 m
Horn	2706	2740 m	—	38 m extrapolated
Schwarzenstein	2778	2811	—	37 m
Grünau	2882	—	2906	24 m »
Sulzenau	2937	—	2960	23 m
			average	30 m

At a critical examination of the above quoted values of the rise in snow-lines it should be remembered, that this rise can be calculated reliable to some extent only at those glaciers, where the tongue has already followed substantially the glacier retreat.

Moreover the glacier has to melt off from the lower parts not from the upper; it therefore must not show parts which become free off snow in the firnregion. This would be an indication of definite unbalance in this glacier, with the inability of feeding the lower parts sufficiently which are in too low an altitude. When calculating the snow-line according to the chief formula 5 in (5) the result would be a line to low.

The average rise of snow-line is 30 meters from 1950-59. This value corresponds to the value of 90 m from 1920-1950 mentioned before.

The fact, that the surface areas of all observed glaciers have been still reduced from 1950-1959 (by an amount of 0,43% per year) is also an indication for the continuation of the glacier retreat. This value is only slight below that of 1920-1950, which amounted to 0,56% per year.

To realize an appropriate and teeming cooperation with meteorologists it should be attempted to include into the observations gauges for ablation and accumulation and also to execute measurements of the ice velocity. The necessary expenditure for measurements and observations will then increase. It should be aspired to solve this extended problem with the help of the meteorologists. — To achieve a still closer connection with meteorology, it would be desirable to chose for the examination of glacier fluctuations and variations in climate typical glaciers in such a manner, that they are situated not too far from permanent meteorological observation stations.

4. THE SELECTION OF TYPICAL GLACIERS

For the investigations of glacier fluctuations glaciers of only few km² area proved to be most favourable in the Eastern Alps. They are influenced quickly by actual climate effects and react even distinctively on the abnormalities in the weather-conditions of one year. Long glaciers follow the effects of changed climate conditions only with substantial retardation and are not influenced by short lasting climate fluctuations to a certain extent. To possibly incorporate short as well as long lasting climate fluctuations, it will be necessary to examine both quick and slow reacting glaciers in international cooperation. To close, an example may be given of two glaciers: one reacting especially quick, one reacting especially slow.

4.1. *The glacier of Bossons*

The glacier of Bossons at the Northern part of Montblanc near Chamonix may be regarded as a particularly quick reacting glacier. As reported in (?) it has a great inclination and a high velocity of ice of 300-500m/year. The latter was measured on the occasion of the Symposium at Chamonix. It should be one of the most rapid glaciers of the Alps. The large level difference of 3400 m from the highest point down to the tongue at a horizontal distance of appr. 7,5 km is passed by glacier waves in a very few years. Obviously such a glacier wave has been effective in the last years and has caused an advance of the tongue by 300 m. By the map 1:10000 of the Montblanc Area of the Institut Geographique Nationale (plotted by an aerial survey) and terrestrial photogrammetric survey in 1958 the advance could be determined.

4.2. *The Glacier of Greenland, nearly 1 million km² large*

It is preponderantly flowing towards West with a velocity generally unknown. The time, in which it reaches the edge from the center, covers many 100 years and connects the present time with the ice-age. By exact surveys in positions and heights the International Glaciological Greenland Expedition (EGIG) has given a base for the examination of changes in ice topography and velocity on the surface. W. Hofmann and H. Mälzer are reporting in this volume on the successfully executed geodetic

surveys, which comprised a West-East traverse of 800 km length through Greenland and a North-South profile of 200 km length near the Western coast.

For the worldwide investigation of fluctuations of glaciers and climate naturally it is required to select systematically further typical glaciers and to measure them possibly with regularity. Then at least in the times of the Geophysical Years numerical figures on the behaviour of glaciers would be available. The above preliminary published examinations serve in principle also the last Geophysical Year and shall lay down and mark for this purpose typical glaciers of the Eastern in their behaviour.

The final publication shall comprise the years 1950-1960. It was and is executed with the assistance of the Alpenverein and the Deutsche Forschungsgemeinschaft and will presumably come out in the «Zeitschrift für Gletscherkunde».

5. FINAL REMARKS

The base for the measurement of glacier fluctuations are in most cases exact contour-lines of the glaciers, surveyed and plotted photogrammetrically at certain timeintervals. One can use terrestrial or aerial photogrammetry as the author has referred at the Toronto Assembly 1957⁽¹⁰⁾. Meanwhile the Symposium of Chamonix has given his suggestion to collect aerial photos of glacier areas. Such photos taken at various dates are without doubt of value in order to compare the stands of glaciers at different times. But one must observe a certain program. Distinct glaciers have to be selected to obtain material not too extensive. Furthermore it is to say, that only photos taken with precise cameras give the possibility to work out datas accurate enough about the height of surfaces of glaciers, about the snow-line, and other essential values treated above. Such values may be needed by the meteorologists studying change of climate of the earth.—Perhaps the planned Symposium about glaciers and climate can suggest directions how each country with glaciers can provide datas in a manner that the world data centers can treat them for scientific use.

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MAIN FEATURES OF THE MODERN GLACIATION OF THE URALS

L.D. DOLGUSHIN (U.S.S.R.)

Institute of Geography of the Academy of Sciences of the U.S.S.R.

SUMMARY

Over 60 glaciers have been established on the Urals, mostly corrie glaciers, with a total area of about 16 km². The glaciers are concentrated in two main regions: Polar Urals (between the 67°30' and 68°09' North) and Subpolar Urals (64°40' - 65°20'). The height of the ends of glaciers varies from 500 to 1300 m over sea level, regularly getting lower from South to North and from East to West.

Despite its small sizes, recent glaciation in the Urals is of great interest. This is a glaciation of quite a peculiar type. The glaciers here are formed and functioning much below the theoretical snow line, which is not reached by a single top of the Ural mountains. They owe their existence to drifting snow driven by the wind into hollow relief forms of leeward slopes. That is why they are all found on slopes and in corries, facing the East and North-East.

Abundant alimentation in winter and intense ablation during the summer determine a considerable glaciation energy, despite the miniature sizes of the glaciers. The movement of ice in typically corrie glaciers of the Ural type differs from the movement of valley glaciers. Deep parts of corrie glaciers move faster than the surface parts, which results in a peculiar shape of the corrie, with an overdeepened bottom and rock threshold. This is facilitated also by the presence of zero temperatures and water in a liquid phase on the boundary of the glacier and its bed.

Recent glaciers of the Urals are a relict phase of young postxerothermic corrie-valley glaciation, separated from the Great Quaternary glaciation by a period with a drier and warmer climate, than the modern one. At the present time the glaciers of the Urals are in a stage of regression.

RÉSUMÉ

Par les investigations des temps récents, l'on a révélé sur les monts Ourals plus de 60 glaciers à superficie totale près de 16 km². Les glaciers sont concentrés dans deux régions principales: sur l'Oural Polaire (entre 67°30' et 68°09' de lat. N) et sur l'Oural Prépolaire (64°40' — 65°20' — lat. N). L'altitude des extrémités des glaciers oscille de 500 à 1300 m au-dessus du niveau de la mer, en s'abaissant régulièrement du sud au nord et de l'ouest à l'est.

Malgré ces petites proportions, la glaciation actuelle des Monts Ouraliens présente grand intérêt — c'est une glaciation de type spécial, ayant lieu beaucoup plus bas que la frontière de neige théorique, qui n'est point atteinte par aucun des sommets de l'Oural. Selon leur origine, les glaciers des monts Oural sont des formations apportées par les vents, survenues et existantes grâce à l'apport de neige par les tempêtes dans les cavités du relief des pentes sous les vents. En conformité à cela, elles sont toutes situées sur les pentes et les kars d'exposition est et nord-est.

Une alimentation abondante et une ablation considérable, au cours de la saison chaude, conditionne une grande énergie de glaciation malgré les dimensions minuscules des glaciers. Le mouvement de la glace dans les glaciers de kar du type ouralien se distingue par son caractère du mouvement des glaciers de vallée. Les parties profondes des glaciers de kar se meuvent plus rapidement que celles de surface, en résultat de quoi la forme du kar est originale.

Les glaciers actuels ne sont point un reliquat de glaciation ancienne; ils sont séparés d'elle par une période xerothermique sèche et plus chaude que la période actuelle. A présent les glaciers ouraliens reculent et certains d'entre eux se trouvent à la limite de leur existence.

Present-day Urals glaciation is of great scientific interest as a special type of glaciation in a mountain country situated below the theoretical snowline.

For a long time by virtue of the continental climate and the low mountains it was considered that there neither are nor can be glaciers in the Urals. The first report about the discovery of glaciers in the Polar Urals was published about half a century

ago (Kertcell, 1911), but it did not attract the attention of glaciologists. It was not until 1929, when cirrus glaciers were discovered in the Sablya Range (Alyoshkov, 1930, 1931), that interest began to be shown for the modern glaciation of the Urals. The Urals glaciers were investigated during the Second International Polar Year (Alyoshkov, 1934, 1935; Boch, 1935). Sixteen glaciers with an area of three square kilometres were discovered in the Urals by that time. Later, the number of known glaciers gradually increased (Govorukhin, 1940; Sofronov, 1945; Khabakov, 1945). Various adjectives were applied to them—"relic", "embryonal", "rudimentary", emphasizing their glaciological defectiveness. Observing the discrepancy between the elevation of the Urals glaciers and the climatic snowline C.V. Kalesnik (1937) was doubtful if they could at all be ranked among real glaciers, qualifying them as "climatically unjustified" and attributing their preservation chiefly to their position in the shade in the depths of the cirques. The author, who in 1945 discovered a new seat of modern glaciation in the Subpolar Urals and visited the earlier known glaciers, has come to a fundamentally different conclusion. Briefly this conclusion is that the glaciers in the Urals exist not despite but thanks to modern conditions and climate and the relief. Nourished by wind-transported snow and avalanches and, as a consequence of this, situated below the climatic snowline, they represent a special "vital form" of modern mountain glaciation. Most of the Urals glaciers have an active rate of movement and carry out extensive geological work (Dolgushin, 1949, 1951). Further investigation confirmed the correctness of this conclusion, while the discovery in the Polar Urals of new glaciers that are bigger than the earlier known ones conclusively dispersed all doubts as to whether the Urals can be ranked as a modern mountain-glacier area. In 1953, an expedition from the Institute of Geography of the Academy of Sciences of the U.S.S.R., headed by the author, brought to light and partially investigated more than 15 glaciers in the northern part of Polar Urals, including the Institute of Geography glacier, which is four times bigger than the Hofman Glacier that was considered the biggest in the Urals, and the Moscow State University glacier, which is 2.2 kilometres long. A hydrological party from the Institute of Geography of the Academy of Sciences of the U.S.S.R. (Dolgushin, Kemmerikh, 1957) discovered another centre of modern glaciation in the Oche-nyrd Range in 1957.

Cirque glaciers situated considerably to the south of the earlier known ones in the Telpos-iz Range were discovered in 1956 and 1959. These were the Govorukhin glacier (Gorbachov, 1958) and the Uzhny Glacier (Kemmerikh, oral report). The latter is indeed the southernmost of the Urals glaciers (63°54' north latitude, 59°10' east longitude).

The northernmost of the Urals glaciers was discovered by the author in 1958 during a study of air photographs and was named for A. N. Alyoshkov. The Alyoshkov Glacier lies in the cirque of the eastern slope of Mount Lyadgei (68°09' north latitude, 65°55'20" east longitude). The region of the occurrence of modern glaciers in the Urals thus meridionally occupies more than 4° latitude. Within these boundaries air photographs showed a few more glaciers, which we have not mentioned in earlier reports: three glaciers in the basin of the Chanshor River (Berg, Kovalsky, Pemyokin), one in the source of the Bolshaya Usa River (Usinsky) and three in the Polar Urals—in the cirques of the Wangyr-Patoksky massif (Patoksky and Aveych) and in the source of the Roshcha-vozh River (Sirin). In addition, a number of large firn fields have been observed in various places.

The Institute of Geography of the Academy of Sciences of the U.S.S.R. set up a glaciological base in the Polar Urals during the International Geophysical Year. It is located on the eastern shore of Lake Bolshoi Khadata-Yugan-lor, at the source of the Khadata River. Stationary observations were conducted on the Institute of Geography, Obruchev and Moscow State University glaciers in 1958 and 1959



Fig. 1 — Distribution scheme of glaciers in Polar Urals; 1- glaciers, 2- rivers and lakes, 3- Principal watershed.

and are continuing in 1960. Base personnel have also conducted route investigations in the Polar Urals. The data of the stationary observations have not yet been systematized, but partially use has been made of some general results kindly placed at our disposal by L.S. Troitsky V.G. Khodakov, I.U. Lebedeva and E.N. Tsykin.

To date we know of 66 glaciers in the Urals, occupying an area of about 16 square kilometres. Their location is indicated on the appended charts (Fig. 1 and 2) and a list is given in Table 1. In addition, in the northern regions of the Urals there are small but numerous young snowfields and migratory snowfields. Avalanche cones

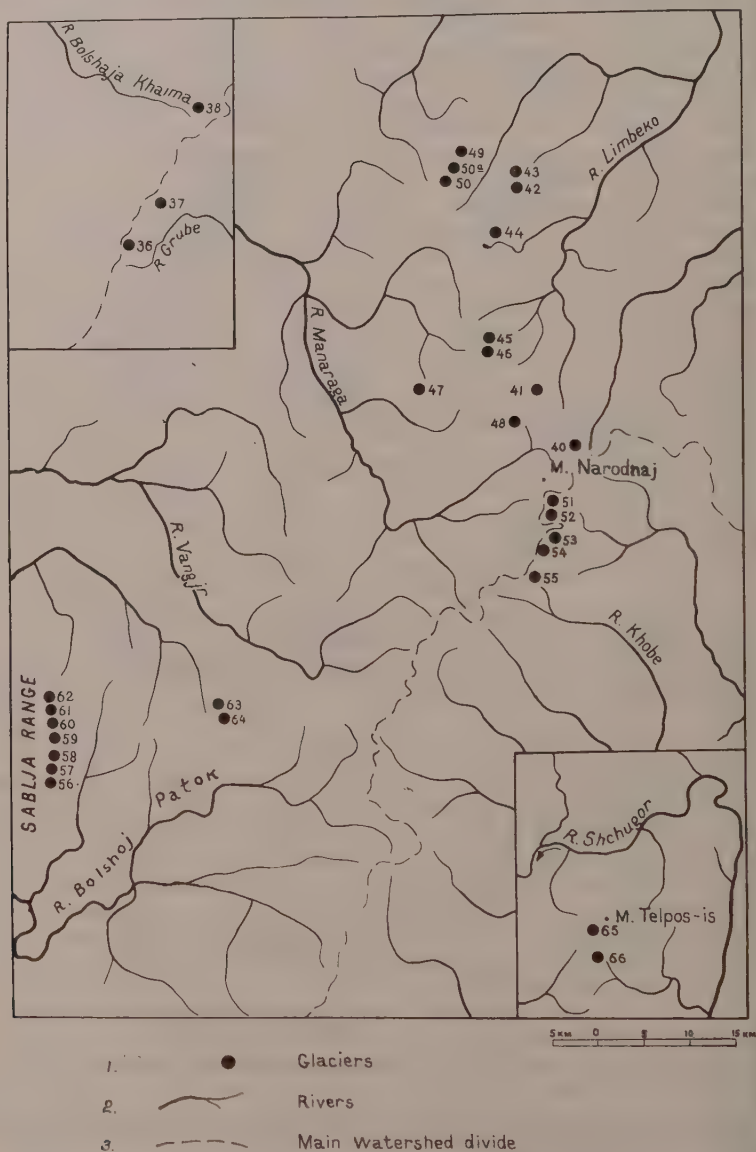


Fig. 2 — Distribution scheme of glaciers in sub-polar Urals; 1- glaciers, 2- rivers, 3- principal watershed.

TABLE 1

Urals Glaciers

No. on dia- gram	Name	Height of ends of the glaciers above sea level (metres)	Length (km)	Area (km ²)	Exposi- tion	Type
1	2	3	4	5	5	7
1.	Alyoshkov	700	1.1	0.50	E	cirque
2.	Terentyev	580	1.0	0.22	ESE	cirque
3.	IGY	650	1.2	0.54	E	cirque
4.	IG	700	0.8	0.40	NE	cirque
5.	Dolgushin	750	1.5	1.00	ENE	cirque
6.	Boch	680	0.4	0.20	E	cirque-suspension
7.	Tushinsky	650	0.3	0.15	ENE	cirque-suspension
8.	Parus	750	0.3	0.05	N	cirque-suspension
9.	Shuchyi	700	1.2	0.80	ESE	cirque
10.	Tronov	650	0.7	0.20	ENE	flat-valley
11.	Plagov	740	0.5	0.16	ESE	leaning
12.	Synok	700	0.5	0.20	E	cirque
13.	MGU	650	2.2	1.10	E	cirque-valley
14.	Karsky	650	1.4	0.65	NNE	cirque
15.	Markov	820	0.65	0.20	E	cirque
16.	Malysh	850	0.4	0.12	E	cirque-suspension
17.	Fyodorov	650	0.75	0.20	NNE	flat-valley
18.	Kalesnik	750	?	0.15	E	cirque
19.	Berg	400	?	0.25	NE	cirque
20.	Kovalsky	690	?	0.25	N	cirque
21.	Lepyokhin	750	?	0.05	N	cirque
22.	Leaning	620	0.3	0.10	ESE	leaning
23.	Rogaty	600	1.5	0.22	NE	flat-valley
24.	Chernov	530	0.8	0.20	ESE	cirque
25.	Obruchev	400	1.2	0.35	E	cirque
26.	Shumsky	570	0.75	0.17	ENE	cirque
27.	Avsyuk	750	0.5	0.11	ESE	cirque
28.	Anuchin	600	0.45	0.08	E	flat-valley
29.	Stantsion- ny	600	0.35	0.08	E	leaning
30.	Oleni	650	0.35	0.09	SE	flat-valley
31.	Institute of geography	800	1.8	1.2	ENE	cirque-flat-valley
32.	Ochkovy	750	0.15	0.12	E	leaning
33.	Usinsky	720	1.5	0.70	SE	flat-valley
34.	Malenky	?	?	0.05	?	cirque

TABLE 1 (continued)

No. on dia- gram	Name	Height of ends of the glaciers above sea level (metres)	Length (km)	Area (km ²)	Exposi- tion	Type
1	2	3	4	5	5	7
35.	Sofronov	?	?	0.10	?	cirque
36.	Varsano- fyeva	850	0.3	0.03	E	cirque
37.	Komarov	850	?	0.05	ENE	cirque
38.	Gorodkov	950	0.3	0.06	E	cirque-suspension
39.	Maldy	1,350	0.3	0.23	E	leaning
40.	Bolban	1,200	?	0.05	ENE	cirque
41.	Limbeko	1,200	0.3	0.13	NE	leaning
42.	Khambal	900	0.3	0.13	NE	cirque
43.	Sirin	950	0.32	0.20	NNE	leaning
44.	Grigoryev	950	0.5	0.16	ESE	cirque
45.	Voeikov	1,000	0.7	0.25	ENE	cirque-suspension
46.	Rikhter	1,200	0.6	0.25	E	cirque
47.	Borzov	1,000	?	0.15	NE	cirque
48.	Manaraga	1,160	0.55	0.21	ESE	cirque
49.	Konus	900	0.25	0.10	NE	leaning
50.	Sale	850	0.3	0.15	NE	leaning
51.	MPY II	1,150	?	0.20	ENE	cirque
52.	Ugra	1,000	0.55	0.21	E	cirque
53.	Mansi	1,200	0.7	0.40	E	cirque
54.	Somnitelny	1,350	?	0.03	SE	cirque-suspension
55.	Khobe	1,100	0.65	0.20	ESE	cirque
56.	Firn 1	750	0.45	0.18	NE	cirque-suspension
57.	Firn 2	800	0.3	0.14	E	cirque-suspension
58.	Firn 3	650	0.6	0.25	S	cirque
59.	Hofman	600	1.0	0.37	NE	cirque with tongues
60.	Firn 5	650	0.35	0.07	NE	cirque-suspension
61.	Firn 6	800	0.2	0.12	E	cirque-suspension
62.	Firn 7	750	0.9	0.25	E	cirque
63.	Patoksky	850	?	0.10	E	cirque
64.	Sosedniz	800	?	0.05	E	cirque
65.	Govoruk- hin	1,060	?	0.18	E	cirque
66.	Uzhny	820	0.5	0.20	S	cirque
Total	area:			16.31		

consisting of ice and firn and covered at the top with products of the destruction of the slopes occur in many cirques. On the bed of the cirques beneath the accumulation of Young moraine, there frequently are sections of dead ice. Permafrost is widespread on the flat surfaces of the highland terraces and plateaux. Subsoil ice occurs in strata and veins in broad ancient glacial valleys where the products of weathering accumulates in tiny fragments. All this shows that there are "on the brink of glaciation" conditions on a vast territory in the northern part of the Urals. This brink has, however, been passed in relatively few places, chiefly in the northern part of the Polar Urals (between $67^{\circ}30'$ and $68^{\circ}09'$ north latitude) and in the Subpolar Urals (between $64^{\circ}40'$ and $65^{\circ}20'$ north latitude), where most of the Urals glaciers are concentrated. Both these areas are elevated and extended sections of the Urals Mountain System, with a complex, very broken relief. The relief has retained traces of a recent greater mountain glaciation in the shape of numerous bowls and cirques, glacier-smoothed gorges and "curly cliffs", crags, terminal and lateral moraines, sections of hilly moraine relief and glacial lakes. The mountains of the Subpolar Urals are several hundred metres higher than those of the Polar Urals and are more broken. However, there is greater glaciation in the Polar Urals. The reason for this is that the Polar Urals seat of modern glaciation is approximately 3° latitude more to the north than the Subpolar Urals. There is also a more noticeable difference in the elevation of the ends of the glaciers in these two areas: in the Polar Urals, glaciers end at an elevation of 400-800 metres (an average altitude of 700 metres), while in the Subpolar Urals they end at an elevation of 600-1,350 metres (an average altitude of 1,050 metres). In addition there is a sharp difference in the height of the glaciers depending on local orographic conditions and their position with regard to the dominating moisture-bringing westerly winds. For example, in the Sablya Range, which is the western outpost of the Subpolar Urals, the glaciers end at an elevation of 600-800 metres, while in the heart of the Subpolar Urals, in the region of Narodny, they do not descend below 1,000-1,200 metres.

Due to the deeper erosional cut of the Polar Urals' valleys with regard to sea level, there is no noticeable difference in the degree the relief of the reviewed parts of the Urals is broken despite the difference between them in the absolute elevation of the mountain peaks and ranges. Although glaciers of the Polar Urals are on the average located 300-350 metres lower than those of the Subpolar Urals, the depth of the cirques holding the glaciers in both areas is approximately the same (350-400 metres on the average in the Polar, and 300-350 metres on the average in the Subpolar Urals, with the extremes being from 200 to 1,000 metres).

There is a distinct law governing the distribution of the glaciers along the macroslopes of the Urals and the direction of exposure (*Table 2*). As that table shows, two-thirds of the glaciers are situated westwards from the main watershed and a third eastward from it. The reason that the modern glaciers are predominantly situated on the West macroslope of the Urals is that it receives much more precipitation, particularly solid precipitation, than the East slope. At the same time, not a single glacier in the Urals is situated on a westwardly exposed slope, facing moisture-bearing currents. All of them lie on the opposite slopes (*Fig. 3*). The reason for this is the predominance of constant high-velocity winds of the western quarter of the horizon, particularly during the cold part of the year. These winds not only bring abundant snowfalls and blizzards but also extensively redistribute the precipitated snow and accumulate it in the hollows of the relief of the leeward slopes. Optimal conditions are created in some of the cirques facing east, north-east and south-east, where most of the Urals glaciers nestle. They are nourished not only by normal precipitation but also by windblown and avalanche-carried snow. In this connection, the areas where snow nourishing the glaciers accumulates are bigger than the glaciers themselves and their morphologically-expressed firn basins. It is only due to this

TABLE 2

Distribution of the Urals Glaciers In Accordance with Slopes and Exposition.

Slopes of the Urals	Exposition of Glaciers								Distribution of glaciers on the macroslopes of the Urals	
	S	SE	ESE	E	ENE	NE	NNE	N	Number	%
West	2	2	5	13	2	13	4	3	44	68%
East	0	2	4	11	5	0	0	0	22	34%
Total: Number	2	4	9	24	7	13	4	3	66	100%
%	3	6	13.5	36	10.5	19.5	6	4.5		100%

that they exist many hundreds of metres below the so-called climatic snowline, which is now not reached by any of the Urals peaks. This unique feature of the avalanche and wind-blown fed glaciers of the Urals gives us grounds for putting them down as a special type of modern mountain glaciation. There are all the more grounds for this because lately "Urals type" glaciers have been discovered not only in the Urals but also in other upland areas (the Altai, Eastern Siberia, the Caucasus).

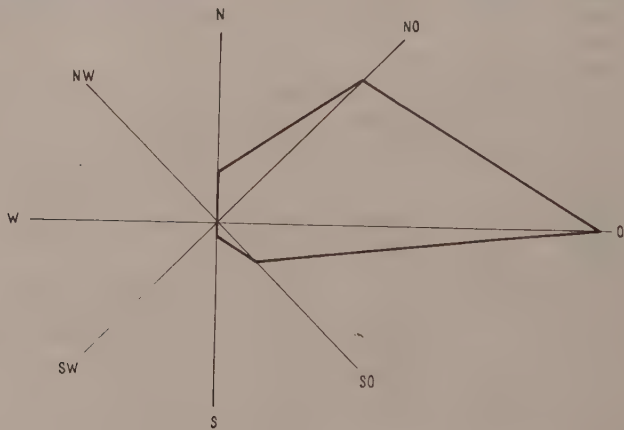


Fig. 3 — Rose of the Urals glaciers exposition.

Modern Urals glaciers may, in conformity with a series of indications, be divided into two main groups. The first group embraces cirque, cirque-valley and cirque-suspension glaciers. The second group comprises so-called "leaning" and "flat-valley" glaciers.

The glaciers of the cirque group from the basis of modern glaciation in the Urals. The biggest of these have tongues extending beyond the boundaries of the cirques into valleys. To distinguish them from the typical cirque glaciers we call them cirque-valley glaciers. Of this group is the Moscow State University glacier (Fig. 4). In spite



Fig. 4 — Plan of the «Moscow University» kar-valley glacier: 1- glaciers' contour, 2- Terminal and lateral moraines, 3- marginal fissures, 4- forn line (middle August, 1953), 5- ice-wells, 6- brooks and gullies on the glacier, 7- horizontal lines (traced after 10 m), 8- sandr field, 9- Cliff scarps.

of their small size, the glaciers of the cirque group have all the basic indications of typical glaciers. Their nourishment is mixed (normal precipitation, wind-borne snow, avalanches). They are formed by: firn (regelation and recrystallization) and glacial ice (infiltration, infiltration-recrystallization, secondary recrystallization). The glaciers are many tens of metres thick (presumably up to 100 metres and over), they move actively and carry on extensive geological work. Historically these glaciers are the modern phase of a recent and vaster cirque-valley glaciation of the Urals, which has left striking traces of its existence.

Leaning and flat-valley glaciers are wind-formed through the accumulation of snowdrifts in the wind shadow on weakly broken terraces of the leeward slopes and in hollows. They comprise strata of firn, and of infiltration and infiltration-recrystallization ice. There are no clear-cut signs of active movement, but the very presence of snow-ice lenses on the slopes facilitates a noticeable activation of the processes of weathering and denudation. As a consequence of this, the glaciers gradually "eat into" the slope, forming a levelled recess. As the glacier grows and cuts into the slope, the conditions for the accumulation of snow improve and in the end the immobile accumulation of firn and ice may turn into an actively moving glacier, and its receptacle—into a cirque (in the event, it goes without saying, that this is not prevented by changes in the external conditions). Leaning and flat-valley glaciers are, in the full sense of the word, modern formations that are now taking shape in favourable conditions of the accumulation of snowdrifts and are the link between glaciers and snowdrifts. As distinct from the cirque glaciers, they do not have inherited receptacles,

frequently developing on quite straight slopes. At the lower edges of the leaning glaciers there is sometimes a pseudo-moraine accumulation of the products of the weathering of the slopes, but the main mass of the small-fragment material is carried away by melt water and solifluction.

We shall give some information about the features of the nourishment, the structure and the regime of the glaciers of the cirque group. Abundant solid precipitation, especially over the cirques of the leeward slopes, snowdrifts, formation of hoar-frost on the surface of the glaciers and on the slopes of the cirques, and gravitational drift of snow from the slopes of the cirques to their foot are what cause the considerable accumulation in the cirque glaciers of the Urals. According to data obtained at the Polar-Urals Base of the Institute of Geography of the Academy of Sciences of the U.S.S.R., the water reserve in the snow cover that accumulates in the course of a year was: on the Institute of Geography glacier—1,000 mm in 1958 and 2,300 mm in 1959; on Obruchev glacier—1,700 mm and 2,500 mm respectively; on Avsyuk glacier—1,300 mm and 2,000 mm. The amount of snow accumulated increases from the mouths of the cirques to their western walls from 500 mm to 4,000 mm (Khodakov, 1960). These first results of direct snow surveys on the Urals glaciers have confirmed the earlier suppositions of the author and some other investigators that snow accumulates in large quantities in the leeward cirques and that this changes considerably in space and in time. According to indirect computations and visual observations made by A. N. Alyoshkov in 1935, 15-20 metres of snow accumulates annually in the cirques of the eastern slope of the Sablya Mountains. S. G. Boch (1946) thought this figure was exaggerated and named another —“not more than 10-12 metres of snow a year”. I believe Boch's estimation is close to the real one. If the mean density of the snow is taken as 0.35 gr/cm^3 , the maximum reserve of water in snow 10-12 metres thick will be 3,500-4,200 mm, i.e., a magnitude that tallies with snow surveys made in the Polar Urals and given above (it must be borne in mind that in the Subpolar Urals the climate is much more humid than in the Polar Urals and that here we are dealing not with mean but with maximum magnitudes of the accumulation of snow in the cirques of the Subpolar Urals).

There is also very intensive ablation in the Urals glaciers. The mean ablation on the Institute of Geography glacier in 1959 was 1,330 mm in a layer of water (Troitsky oral report), while for the Moscow State University glacier it was 1,534 mm for 35 days of observation (ablation lasted for a period of 45 days), or 44 mm a day (Lebedeva, 1960). Lebedeva further points out that solar radiation plays a much smaller role in the melting of the glaciers than turbulent heat exchange with the atmosphere and the heat of condensation, which together exceed 54%. This confirms and specifies what I stated in 1949, e.g. that in the northern parts of the Urals, as a consequence of predominating nebulosity and little sunlight, the role played by direct solar radiation in the melting of glaciers and snowfields cannot be very big. Snow and ice melt chiefly through the influence of convection. As they cross the Urals, moist and warm westerly winds bring not only precipitation but also warmth. “Thus, the wind is not only a major factor in the accumulation of snow in the cirques, but also a very important factor contributing towards the melting of snow. In the Urals, glaciers occur hundreds of metres below the climatic snowline not because the melting of snow is not intensive (on the contrary, it is quite considerable), but chiefly because there is very intensive accumulation of snow” (Dolgushin, 1949, 1951). This determines the very energetic glaciation in the Urals compared with similar glaciation in conditions of inconsiderable accumulation and ablation (for example, the Central Asian cirque glaciers).

The sharply expressed seasonal character of the processes of ablation and accumulation makes for the stratified structure of the glaciers, this being reflected on the surface in the form of stripes (on even glaciers there are from several tens to

200 stripes). Annual and old stripes occur in the glaciers. This is excellently seen on air photographs of many glaciers. For example on Manaraga glacier. Boch counted 105 stripes. With the aid of air photographs it is being established that these layers group in packets, of which 13 are seen on Manaraga glacier, whose surface is free of snow and moraine. The projection of each packet on the surface averages about 30 metres in length. The alternation of these packets, we believe, reflect the fluctuation over a period of many years of the conditions of accumulation and ablation that repeat themselves periodically.

Watching the elements of stratification of the firn and ice in the Urals cirque glaciers we observed changes in the bedding of the layers from parallel to the surface in the region of accumulation to vertical in the terminals of the glaciers. This phenomenon may be explained only by surmising that deep in the cirque glaciers of the Urals the rate of movement of the ice is much greater than on the surface. This supposition is all the more probable because it alone explains the form of the cirques with their characteristic very deep bottoms and rocky gorges at the mouth. The rate of movement was measured on very few glaciers and then only on the surface. In the period from August 1958 to August 1959, the maximum surface velocity of the Institute of Geography glacier was about 5 metres (Troitsky, oral report). Observations of the velocity of movement of Mansi glacier showed that it was not less than 5 metres a year taken as a mean for 12 years (from 1933 to 1945). The velocity of 12-15 metres a year indicated by me was obtained indirectly and is, possibly, somewhat exaggerated. Deep in the cirque glaciers the velocity must be greater than on the surface. Despite the small absolute values for the velocity, "metabolism" in the cirque glaciers of the Urals is rapid thanks to the small horizontal dimensions. They carry out extensive geological work, intensifying the weathering of the rock and transporting the products of weathering to the mouth of the cirque, where these products are taken over by melt and rain water. The glaciers also continually deepen their bed. Most of the cirque and cirque-valley glaciers of the Urals have lateral and terminal moraines.

According to their temperature regime they are warm glaciers. According to observations made by E. N. Tsykin, negative temperatures in the body of the Institute of Geography glacier are observed only on the surface levels and are from -1.2°C to -1.7°C . at a depth of 10 metres and from -0.4°C to -0.7°C at a depth of 25 metres. Below 45-50 metres the temperature of the ice must be about zero (Tsykin, 1960). Thus is justified the supposition that at the bottom of the glaciers there are "water smears" that facilitate movement.

In the lifetime of the Urals glaciers, as we have already pointed out, there were frequent changes of periods of worsening and improving conditions. At the present time the short fluctuations of these conditions take place from year to year against the background of a general retreat of the glaciers. This is testified to by the high location of modern lateral and terminal moraine above the surface of the glaciers, the formation of hollows and lakes on the glacial tongues above the terminal moraine, the isolation of the terminal moraine swells from the body of the glaciers, the conversion of sections of the tongues into dead ice and their destruction. On Moscow State University glacier, the lateral moraine, frequently with an ice core, is 40-50 metres above the surface of the glacier and continues to grow. The end of the glacier is badly broken and is crumbling. High lateral moraine with an ice core is also typical for the Institute of Geography glacier, which likewise has a broad belt of terminal moraine formations. Lying between the lateral and terminal moraines the tongue of the glacier has acquired the shape of a bed with a lake. Comparing large scale air photographs of the Institute of Geography glacier taken in 1958 with was observed in 1953, it may be noted that there have been substantial changes on its surface: the prevalence of lateral moraine over the surface of the ice has noticeably increased, the pollution of the glacier with moraine has become more considerable, the lake on the glacier

has dropped and greatly diminished in size, the uncovering of moraine-saturated ice rising up to 10-15 metres above the lower end of the lake has dispersed and levelled out. The glaciers of the Subpolar Urals that I have visited in 1945 and later (Yurga, Mansi, Moscow State University II and others) are now much more polluted with moraine material and the steep precipice of Mansi glacier facing the lake, which was more than 20 metres high, has levelled out. The general snow cover of the Subpolar and Polar Urals has also diminished. Many old snowfields have melted completely, others have greatly diminished in size. In this respect it is interesting to compare the modern state with what prevailed when these parts of the Urals were investigated earlier. While belying the presence of glaciers in the Urals, E. Hofman at the same time wrote in his report about his expedition to the Northern Urals in the middle of the past century that great layers of ice lie in many of the Urals valleys and ravines in the course of the whole summer. Hofman possibly did not see the glaciers because they were concealed by a thick layer of snow. Thus, passing the eastern foothills of the Sablya Mountains, he wrote: "its 14 jagged peaks were not covered with snow, but neither did snow blanket the rest of it" (Hofman, 1856, p. 182). That is precisely where Alyoshkov discovered the first Urals glaciers 80 years later. At the close of the summer of 1850 (August 2), Hofman wrote: "There are big snowfields at the western foothills of the Polar Urals, in tundra protected against the sun. They did not grow this summer, which Andrei (the guide) called hot and dry. Larger masses of snow are to be found in the Urals valleys and there are, I am told, huge never melting masses of it on the other side of Mount Lyadei (*ibid*, p. 141)". It is interesting to note that the Alyoshkov glacier, the northernmost of the cirque glaciers known in the Urals, was discovered in this very spot in 1958 with the aid of air photographs.

In 1909, Baklund, who had his camp where the base of the Institute of Geography of the Academy of Sciences is now located, wrote: "the snow cupola at the top of the Kholong, or, to be more exact, the more clear-cut walls of the cirque were named Khard-Naurdy-keu by foreigners" (Baklund, 1911). There can be no doubt that he meant the Institute of Geography glacier, which he mistook for a snow cupola. Further he wrote: "The view that opened to the west amazed me for its alpine beauty. On either side of the lake (Bolshoi Khadata-Yugan-lor) there is a whole row of conical peaks, steeply descending towards it. Some of the peaks are separated from each other by V-shaped hollows, the beds of which lie higher above the level of the Khadata River. A number of cirques and recesses, some facing the lake, others the valley of the Khadata and lying buried beneath deep snow, complicate the relief still more" (*ibid*, p. 52). Traces of this broad development of snowfields in the Urals are also well-expressed in air photographs.

In the Polar and, especially, in the Subpolar Urals there are numerous signs of a vaster mountain glaciation than today and also traces of the retreat of the glaciers to their present size. We can define two stages of this glaciation, which did not go beyond the framework of the axis, more elevated part of the Urals. The first stage (the maximum) is fixed by a complex of moraine accumulations in the mountain valleys at a distance of 10-15 kilometres from their sources. Most of the glaciers were much smaller and among them there were many cirque and cirque-valley glaciers. They came down the Subpolar Urals to 400-700 metres and the Polar Urals to 300 metres (in places) from the mark. The second stage is observed in the mounths of very many cirques occupied or unoccupied by glaciers. Some of the glaciers developed short tongues into the valleys (up to 3-5 kilometres long). Traces of this stage are clearly expressed and show a direct link with modern glaciation.

The mountain glaciation of the Urals is not a direct continuation of the ancient sheet glaciation, when the glacial sheet completely covered the territory of the Polar and Subpolar Urals, with the exception of separate nunataks. Between them there was an interval, during which predominance was taken by the processes of erosion and traces of the ancient sheet glaciers in the mountains were greatly obliterated.

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MAIN PARTICULARITIES OF GLACIATION OF CENTRAL ASIA ACCORDING TO THE LATEST DATA

L.D. DOLGUSHIN (U.S.S.R.)

Institute of Geography of the Academy of Sciences of the U.S.S.R. Moscow

SUMMARY

During 1958-1959 a joint Expedition of the USSR Academy of Sciences and the Academy of Sciences of the Chinese People's Republic had inspected glaciers in Alpine mountain ranges of Nan Shan and Eastern Tien Shan. In Nan Shan 965 glaciers have been recorded, mostly of a corrie and corrie-valley type, occupying an area of 1280 km². The glaciated territory of Eastern Tien Shan is about 2000 km²; we find there glaciers of a valley type up to 30 km long and with an area up to 100 km².

The territories studied belong to the Central-Asiatic continental glacier province, with an extreme continental climate. Their glaciation is substantially different from glaciations of adjoining territories of Western Tien Shan, belonging to the subcontinental Atlantic glacier province and of Eastern Nan Shan located on the margin of the subcontinental Pacific glacier province. These differences are reflected in the height of the snow line, amount of supplied and discharged matter of the glaciers, the volume of glaciation energy, the temperature and hydrological regime of the glaciers. Thus, the glaciers of Western and Central Nan Shan, as well as those of Kuenlun, end at much greater heights than the glaciers of Tien Shan and Eastern Nan Shan; they are characterized by a less intense exchange of matter in the glaciers, a lesser energy of glaciation, a lower temperature regime, prevalence of a surface run-off of meltwater, a more inert reaction to a change in external climatic factors.

Numerous traces of ancient glaciation have been discovered. Old glaciers were descending in Nan Shan up to the marks of 2900-3800 m. They came out to the bottoms of intermontane depressions, but were not filling them, as it was thought before. In Tien Shan ancient glaciation was more extensive. Largest glaciers descended up to 1800 m. At the present time Tien Shan glaciers retreat faster than the glaciers of Nan Shan.

RÉSUMÉ

L'expédition de l'Académie des Sciences de l'U.R.S.S. et de l'Académie des Sciences de la République Populaire de Chine en 1958-59 a investigué les régions glaciaires alpines du Nan-Chan et du Thian Chan Oriental. Dans le Nan-Shan, l'on a révélé 965 glaciers couvrant une superficie de 1280 km². L'aire de glaciation du Thian-Chan Oriental excède 1700 km².

Le Nan-Chan Central et Occidental, l'extrémité Orientale du Thian-Chan et le Kun-Lun sont situés dans les confins de la province glaciaire continentale de l'Asie Centrale, et leur glaciation diffère grandement de celle du Nan-Shan Oriental, situé dans la province subcontinentale Pacifique, et du Thian-Chan Central, qui se rapporte à la province subcontinentale de l'Atlantique.

Ces différences se font sentir sur la hauteur de la ligne de neige, sur la valeur de la recette et de la consommation de la neige, sur l'énergie de la glaciation, sur le régime de température et sur le régime hydrologique des glaciers.

Notamment, les glaciers de Nan-Chan se trouvent à une beaucoup plus grande altitude, en comparaison aux glaciers du Thian-Chan (conformément : 4000-5000 m et 2800-3500 m); ils se caractérisent par des plus petites valeurs de recette et de dépense de la substance, par un régime à température plus basse, par une prépondérance de déversement superficiel des eaux de fonte et par une série d'autres indices.

Des indices de l'ancienne glaciation dans le Nan-Chan peuvent être tracés jusqu'aux marques 2900-3800 m. Les glaciers anciens affleuraient sur les fonds des dépressions, mais ne les remplissaient pas, comme on le pensait jadis. Dans le Thian-Chan, la glaciation ancienne était plus vaste : des glaciers immenses descendaient jusqu'aux avant-pays avec des marques 1800-2000 m. Actuellement, les glaciers reculent sur le Thian-Chan plus vite que sur le Nan-Chan. Sur certains glaciers du Nan-Chan, il y a des indices d'avancement actuel.

The areas of mountain glaciation in Central Asia are among the least explored and are accessible with great difficulty. At the same time the ices and snows covering many of the Central Asian ranges and uplands are one of the main sources of water supply for oases at the foot of the mountains.

In connection with the problem of supplying with water the arid north-western provinces of China, a special expedition for the study of the high altitude ices and snow and the working out of methods for their practical utilization was organized jointly by the Academy of Sciences of the Chinese People's Republic and the Academy of Sciences of the U.S.S.R. in 1958-1959. As a result of the field work carried out and the analysis of the data obtained by means of aerial photography, the explorers determined the specific features of glaciation of the Nan Shan and the eastern Tien Shan ranges; determined (as a first approximation) the areas of glaciation and the minimum water resources contained in the glaciers, and carried out experiments in artificially accelerating the thawing of ice and snow in the mountains. The data obtained yield information on the morphology, hydrological regimes, temperature conditions, dynamics and evolution of the glaciers. For the first time in Central Asia, glaciological observations lasting all the year round, have been started at the newly established high-altitude stations: Laihukow in the north-western part of the Nan Shan (36°30' N., 96°30' E, 4060 m above sea level) and Tasikow in the Eastern part of the Tien Shan (43°06', 87°15', 3480 m above sea level). The author of this article who took part in the explorations work on the Soviet side, restricts his report mainly to the new data received by the expedition and presents some general results.

The main characteristics of the present-day glaciation of the Nan Shan, the Eastern Tien Shan and the western Kun Lung are to be found in Table 1.

The data concerning the glaciation of the Nan Shan are somewhat more complete than the others, although in the past the information available on this range was limited to brief descriptions (Przhevalsky, Roborovsky, Kozlov, Obruchev, Stein). Now over 1000 glaciers with a total area of about 1500 km² have been found. The approximate amount of water in these glaciers is close to 50,000,000,000 m³.

The glaciers of the Nan-Shan are numerous, but relatively small in size. Most of them are suspended and tarn glaciers (40 and 30 per cent respectively). Tarn valley and valley glaciers of a small size are widespread, forming 25 per cent of the number of glaciers and 35 per cent of the glaciation area. The glaciers of these types, alongside the tarn glaciers are particularly important from the hydrological point of view. The glaciers originating on flat summits, which are concentrated in the south-western ranges of the Nan Shan, constitute 2,5 per cent of the total number of glaciers and about 10 per cent of the glaciation area, but their role in the run-off is insignificant. The area of perennial snow glaciers and small firn glaciers constitutes approximately 2,5 per cent of the total.

The expedition did not find any extensive glaciation sheet on the Ritter (Chahanbotu) range, which was indicated for the first time on the maps drawn up by Roborovsky and Kozlov under the name of Guchin-gurbu.shahalgyn (Roborovsky, 1900). According to Obruchev (1931), the area of this glacier must be close to 800 km². Here, as well as on the neighbouring ranges (Turgen-daban, Tsaidamo-shan) about thirty small glaciers originating on flat summits have been found. The biggest among them does not exceed 17 km². The total area of all the glaciers originating on the flat summits of the Nan Shan constitutes 120 km², but they are scattered at different spots on the summits of several ranges, and do not form any continuous glaciation sheet.

The foci of glaciation in the Nan Shan territory are distributed with a certain regularity. Glaciation is most considerable in the central and western parts of the territory, where the altitude of the mountain ranges, as a rule, exceeds 5000 m. Glaciation in the south-eastern part of the Nan Shan is insignificant, in spite of a more humid

TABLE 1

The glaciation of Nan Shan, eastern Tien Shan and western Kun Lung.

Ranges (Altitude above sea level in meters)	Number of glaciers	Area of glaciation	Altitude above sea level				Ancient moraines	
			Snow line		Glacier margins		north slope	south slope
			north slope	south slope	north slope	south slope		
1.	2.	3.	4.	5.	6.	7.	8.	9.
NAN SHAN								
(From expedition report published in 1958, and the observations of the author)								
Lengjungling (4500-4800)	116	118,5	4250	4450	3860	4150	—	3300
Richthofen (4500-5900)	271	301,0	4360	4600	3880	4150	3150	—
Tolai-Shan (4500-5000)	79	46,1	4550	—	4200	—	2900	—
Tolai-nan-Shan (4500-5000)	85	70,6	4550	4600	4150	4400	3600	—
Ema-shan (4500-5300)	107	161,7	4700	4750	4250	4450	—	—
Süss (5000-6200)	130	500,0	4650	5100	4350	4650	4000	4200
Humboldt (4500-5500)	100	76,0	4700	4800	4400	4550	—	—
Ritter (4500-5500)	42	63,5	4750	—	4600	—	3900	4130
Mushketov (4500-5500)	102	175,6	4750	5150	4600	4750	—	—
Tsaidamo-shan (5000-5800)	25	49,0	5200	—	4800	—	—	—
Total for the Nan Shan range	1055	1565,0	4645	4780	4300	4440	—	—

TABLE 1 (Continued)

Ranges (Altitude above sea level in meters)	Number of glaciers	Area of glaciation	Altitude above sea level						Ancient moraines	
			Snow line		Glacier margins					
			north slope	south slope	north slope	south slope	north slope	south slope	north slope	south slope
1.	2.	3.	4.	5.	6.	7.	8.	9.		
Kurlyk-tag (4000-4900)	66	102,0	3900	4100	3400	3450	2100	2260		
Barkhalu-tag (4000-4200)	23	16,0	3900	—	3250	—	—	—		
Bogdoshan (5000-5500)	24	70,7	3750	3950	3200	3300	2150	2340		
Irenhabirga (4000-5500)	?	673,0	3800	4050	3200	3550	3070	3400		
Source of river Haidyk-gol	38	40,0	3850	—	3500	—	2950	3200		
Halyk-tau (4000-6000)	?	290,0	3850	4000	3400	3550	—	—		
Kokshaal-tau	?	521,0	3750	4150	2950	2900	—	1900		
Total for the Eastern Tien-Shan	?	1702,7	3830	4050	3270	3350	—	—		
WESTERN KUN LINGI (According to G. Sobolevsky)										
Basin of the rivers Kiliang, Karakash and Yurungkash	68	—	4900	5180	4370	—	3600	4150		

climate,—a phenomenon due to the lower altitude of the mountains. The only important glacier focus in this part of the mountain system is situated on the Lenglungling range, rising to 4800 m above sea level.

In conformity with the aridity of the climate which increases in the western and south-western directions, the level of the snow line and that of the ends of the glaciers becomes increasingly higher in that direction. On the northern slope of the Lenglungling range, the snow line lies at an altitude of 4250 m, while on the same slope of the Tsaidamo-shan range it is situated at 5200 m. The level at which the ends of glaciers are situated rises accordingly from 3800 to 4900 m. The general regularity in the modification of the snow line level is displayed most distinctly in comparing the levels of flat-summit glaciers, on which the influence of the local orographic conditions is least noticeable. For instance, the snow line level of the Hsiaoshalung flat summit glacier is situated at 4500 m (Richthofen range in the eastern part of the Nan Shan) while in the glaciers of the same type situated in the south-western ranges of the Nan-Shan, this level is never below 5150-5200 m. Alongside this general regularity, reflecting the changes in climatic conditions, sharp fluctuations in the height of the firn line of glaciers are observed, which are due to local orographic conditions.

The degree of the glaciation of the mountain slopes is found to be directly dependent on their exposure. The snow boundary on the northern slopes usually lies at a level which is 200 or 300 m below that of the southern slopes. The Tolai shan, the Tsaidamo-shan and the Tolai-nan-shan carry on their northern slopes many glaciers which often descend by many hundred metres below the ridges of the watersheds, while the southern slopes of these ranges are practically free from modern glaciation. On the whole, about 72 per cent of the Nan Shan glaciers lie on the slopes with an northern exposure; 21 per cent of them are situated on the southern slope, and only about 7 per cent fall to the share of the eastern and western slopes (Fig. 1). This is not only due to differences in insolation found on slopes with a different exposure, but is also accounted for by orographic conditions, such as the latitudinal or nearly latitudinal direction taken by the Nan Shan range (a direction of this kind favours the formation of slopes and hollow relief forms on the northern and southern sides).

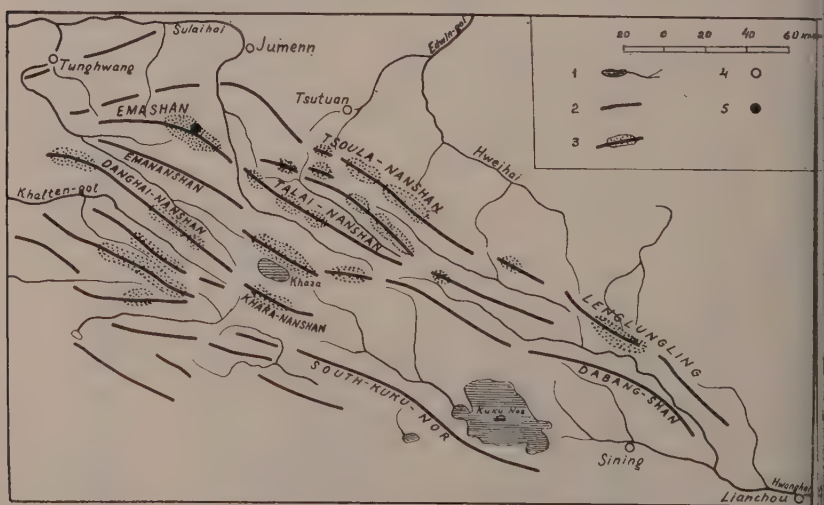


Fig. 1 — Scheme of present day glaciation foci of Nan-Shan: 1- rivers and lakes; 2- mountain ranges, 3- foci of present day glaciation, 4- settlements, 5- glaciological station Laohukou.

preferably to those exposed to the East and West). A considerable role is also played by the greater altitude, width and ramification of the northern slopes of most Nan Shan ranges compared with their southern slopes.

The above mentioned specific features of the geographical distribution of the Nan Shan glaciers are reflected in the relative role of the thaw waters of the glaciers in the run-off of the rivers, which rises from 5 per cent in the South-East to 30-35 per cent in the North-West and to 60-80 per cent in South-West, parallel to the increasing general aridity of the climate in the same direction.

From the morphological viewpoint, most of the Nan Shan glaciers are distinguished by firn basins with steep slopes, a great number of fissures and ice drifts, as well as tongues descending along relatively gentle slopes and reaching the bottoms, of valleys. The surface of glacier tongues has usually a small number of cracks, but is as a rule, distinguished by considerable ablation breaks. The glacier tongues end in steep protruding "fronts" 40-60 m high. In some of the glaciers the ends of the tongues are concealed by a thick cover of moraines. In some spots big terminal moraine ridges up to 100 m high are found. They contain inclusions and pieces of dead ice. Lateral moraines, both ancient and modern, are widely developed. At the same time there are many glaciers with very weakly developed moraines, which practically leave only fluvio-glacial deposits in their retreat. The destruction of the glacier tongues, due to the thermal radiation of the basic slopes, the destructive action of the lateral (marginal) streams of thaw waters and the periodic crumbling of the icy slopes undermined by water,—is more manifest on the sides than in the central part (axis) of the glacier tongue. Due to this, the tongue bulges in its central part, and on its edges vertical ice walls are formed. By studying them it is possible to get an idea on the internal structure of the glaciers. In these sections various forms of ice tectonics folds, breaches, faults, etc.) may be found, which make it possible to judge of the complex nature of glacier motions and their effect on their bed. The alternation of ice layers enriched by moraine inclusions to a different degree bears witness to the repeated succession of the conditions under which a glacier existed, particularly the succession of periods with predominating ablation and those when accumulation became the prevailing characteristic.

The structure of the upper horizons of firn and ice in the firn basins of many glaciers permits the explorers to determine the degree of pure accumulation (minus run-off and evaporation) in a number of preceding years. At the typical valley tarn glacier "July 1", situated on the northern slope of the Tolai-shan range at an altitude of 4850 m, pure accumulation amounted in the last six years to a mean annual figure of 400 mm (in terms of water), for all the ablation zone. As the "July 1" glaciers has a glacier coefficient of 1.15, this ablation must be compensated by accumulating in the firn basin at least 355 mm of precipitation. This figure is close to the mean annual accumulation in the firn basin, established by means of analysis. As during the warm season thawing sets in almost on the entire surface of the Nan Shan glaciers, including the firn basins, the amount of snow precipitation must naturally be more considerable than pure accumulation. We do not know the figure yet, but it should not be very considerable. According to the observations made in 1958-1959, the line of the zero balance at the "July 1" glacier passed at an altitude of 4550 m in the central and western parts of the firn basin, where avalanches play an important role, while in the eastern part of the glacier it was situated at an altitude of 4650 m. These altitudes are close to that of the firn line observed in 1958, towards the end of the ablation period. According to technical observations, the speed of the glacier movement in the central part of the tongue amounted on an average to 11.4 m a year, the glacier being about 70 m thick in this part (according to the Lagally formula). The data cited here indicate that the glaciation energy of the Nan Shan glaciers is less considerable than that of glaciers in many other mountainous regions in the temperate zone.

Another essential feature of the Nan Shan glaciers is their low temperature regimen. For instance, at the "July 1" glacier, at an altitude of 4580 m, the temperature of the ice at a depth of 9 m was $-7^{\circ},9\text{C}$ (taken on July 18, 1959 by means of a thermometer driven into the ice). At the No. 20 glacier of Laohukow in the Ema-shan range at an altitude of 4440 m, the temperature of the ice at the depth of 5 m amounted to $-10^{\circ},1$ (June 31, 1959). In the middle of the warm season, low temperatures below zero were observed already in the ice at a depth of one metre (fig. 2). The first data of temperature observation (on which there are not many, for the time being) show that the Nan Shan glaciers are among the coldest glaciers of the temperate zone (they are colder than the Tien Shan glaciers and their temperature is close to that of the Suntar-Hayat glaciers situated in the neighbourhood of the pole of cold of the northern hemisphere). The low temperature regimen of the Nan Shan glaciers is a consequence of the local sharply continental climate. In the course of the winter season, when cold and very dry weather of the anti-cyclone type prevails, and there is practically no precipitation, the temperature of the glaciers becomes much colder. The summer is cold, with frequent snowfalls, and there are fluctuations above and below zero practically every day. It is natural that under these conditions both ablation processes and the warming up of the glaciers take a very slow course. The presence of temperatures below zero in the layers of ice near its surface lead to a predominance of superficial run-off of the thaw waters and lend some specific features to the processes of ice formation. According to the data of the meteorological stations, about 80 per cent of the annual precipitation in the foothills and some of the mountain valleys of the Nan Shan occurs in the period from May to September. The data taken at

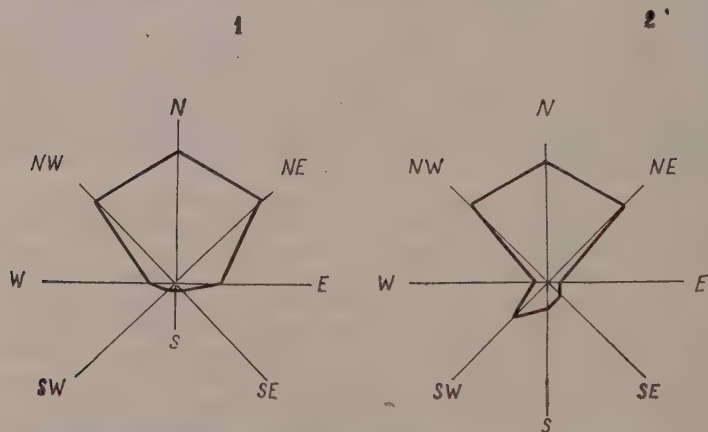


Fig. 2 — Roses of glaciers' exposition (m %): 1- of Western Kun-Lun, 2- of Nan-Shan (for the whole mountain system, without the range Züss).

the Laohukov high altitude station (4060 above sea level), as well as the observations of the expeditions have shown that this ratio also applies to the high altitude zones with the difference that there precipitation usually occurs in solid form (snow, hail, sleet) even during the warm season. Thus the Nan Shan glaciers are mainly fed by the precipitation of the warm season. The process in which snow is transformed into ice takes a very rapid course. In the ablation hours, thaw sets in not only in the glacier tongues, but also in their firn basins. Yet the thawing of the firn basins yields practically no run-off, as the thaw waters penetrate to the level of constant negative temperature, which is located near the surface and freeze again. This is favoured by the drop

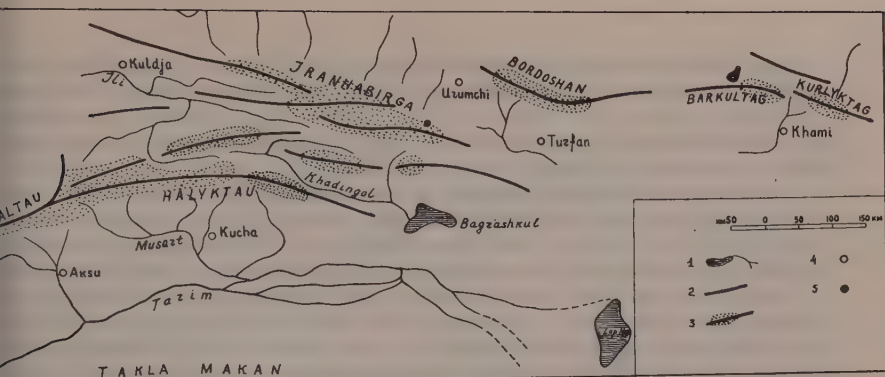


Fig. 3 — Scheme of present-day glaciation foci of Eastern Tien-Shan: 1- rivers and lakes, 2- mountain ranges, 3- foci of present-day glaciation, 4- settlements, 5- Alpine station Dasi-Kou.

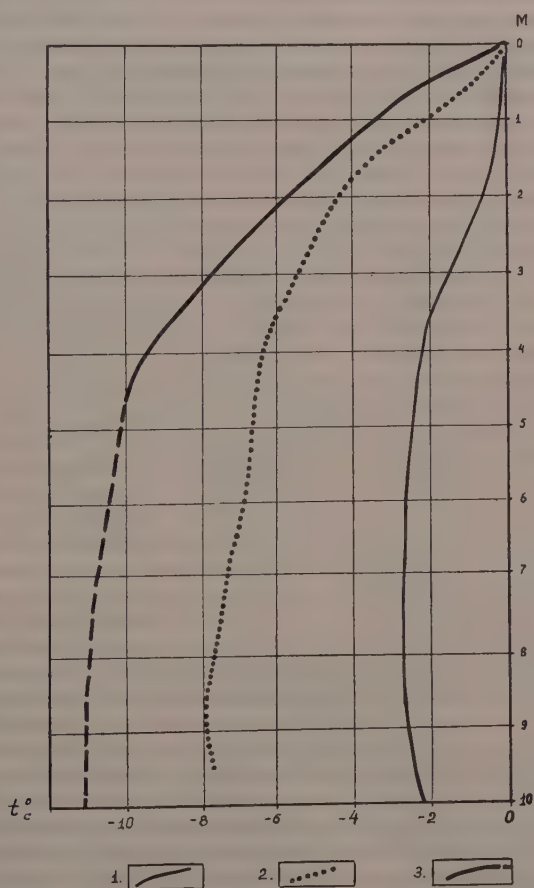


Fig. 4 — Temperature graph in the ice's thickness: 1- of Karabatak glacier in central Tien-Shan, altitude 3460 m (average data for summer periods 1948, 1949 and 1949-50 acc. to G.A. Avsjuk); 2- Glacier «Ist July» in Central Tien-Shan, altitude 4580 m, 18 July 1959, 3- of glacier no. 20 Laohukou in NW Nan-Shan, altitude 4440 m, 31 of June, 1959.

of temperature below zero, which frequently occurs at night. The frequent repetition of this process alternating with snowfalls leads to the formation of a stratified layer consisting of firn and infiltration ice. In the course of one or two years the ice strata replace the firn almost entirely. Therefore the Nan Shan glaciers consist of ice almost entirely, not only on the tongues, but in the firn basins.

The glaciers of Eastern Tien Shan differ considerably from those of Nan Shan. The glaciation area of the Eastern Tien Shan exceeds 1700 km^2 , with an estimated amount of water equalling about $100,000,000,000 \text{ m}^3$. The Tien Shan range forms a complex of ramifications and outcrops and rises to an altitude of 5000-7000 metres. The situation of the Tien Shan mountain system in the sphere of influence of the Atlantic determines the lower level of the snow line and leads (in combination with orographic conditions) to the formation of glaciers distinguished by larger size and belonging to a different type. Apart from the types of glaciers found in the Nan Shan, Tien Shan glaciers are often of the complex valley type, with scores of tributaries and a length of 20-30 km. The area of Karagul, the biggest of these glaciers constitutes 90 km^2 (without the avalanche spots from which it is fed). The tongue of the glaciers are encumbered by moraines over a distance of many kilometres. They abound in fissures, thermo-karst precipices, lakes and other signs pointing to degradation processes. Considerable parts of the tongues have become areas of dead ice. The valley glaciers in the terminal parts of the tongues have no surface run-off of thaw waters: under-ice and intraglaciar run-off predominate. Out of the cave of a glacier a river of considerable size often takes origin. The stream which issues from the cave of Karagul glacier had in July 1959 a mean monthly discharge of 26.5 m^3 per second, the maximum discharge being 50 m^3 per second. The ablation of the glaciers in the Eastern Tien Shan mountains considerably exceeds that of the Nan Shan glaciers, a fact which is accounted for by the extremely low hypsometric position of their tongues with regard to the snow line (the ends of the glaciers Karagul, Muzart and Tugbelich on the south-eastern side of the Khan-Tengri massif descend respectively to 2920, 2730 and 2780 m, the snow line being situated at an altitude of 4150 m). The great negative difference in glaciation is determined by the tremendous development of the gorges and precipices, and, in this connection by the predominance of avalanche reinforcement to the glaciers from the surrounding slopes.

The intensity of glaciation in Eastern Tien Shan drops in the direction from West to East, in accordance with the increasing aridity of the climate in the same direction. The Tien Shan ranges lying in the extreme East of the system (Kurlyk-tag, Barkultag) are closer in type to the western Nan Shan than to the eastern Tien Shan in the glaciation areas, the types of glaciers and their regimen.

As far as it is possible to judge from the scant data published, the glaciation of the northern ranges of the Kun Lun is also closer in its main characteristics to the south-western part of the Nan Shan.

Comparing the glaciation of the inland mountains of Central Asia with that of the surrounding peripheral ranges, we find some essential features of difference between them. These differences find expression in the regular rising level of the snow boundary from the periphery towards the inland parts of Central Asia; the drop in the intensity of ablation and accumulation processes (and consequently the energy of glaciation) in the same direction; in the lower temperature regimen of the Central Asia glaciers; in the predominating surface run-off of thaw waters in the glaciers of the inland mountains of Central Asia, and in the lower regularity of the run-off in the glacier-fed rivers of Central Asia (inland) compared with these of the peripheral ranges. The morphological differences are also essential. All this makes it possible to raise the question of singling out a special Central Asian continental glacier province, including the glaciers of the Nan Shan (with the exception of its south-eastern part which is in the sphere of influence of the Pacific monsoon), those of the Kun Lun, the

Altyn-tag and the ranges of the eastern part of the Tien Shan. We leave open the question of the inner areas of the Tibetan highlands as there are no data on its modern glaciation.

The continental Central Asian glacier province borders in the West on the subcontinental Atlantic province (Central and Western Tien Shan, Pamirs-Altai, Altai, etc.); in the East on the subcontinental Pacific province (Eastern Nan Shan, the eastern borderline ranges of the Tibet); on the South, with the Indian monsoon province (Himalayas, Karakorum).

The modern stage in the evolution of Central Asian glaciers is distinguished by their retreat everywhere. The intensity of this retreat is greatest in the region of Mount Khan-Tengri in the Tien Shan and lowest in the north-western part of the Nan Shan, on Ema-shan range, where stationary and even advancing glaciers are found alongside retreating ones. Yet in the Nan Shan and especially in the Tien Shan, degradation processes obviously predominate. In the last fifty years, for instance, the Muzart glacier (region of Khan-Tengri) has retreated 560 m, according to verbal information and geomorphological indications. In the terminal part of its tongue, it has become thinner by 30 or 40 m over a distance of several kilometres.

Ancient glaciation was considerably more extensive and the ice sheet was thicker than in the present-day process. Traces of ancient glaciation (lateral and terminal moraines and geomorphological signs) are found in the Nan Shan up to an altitude of 2900-3900 m; in the Kun Lun, up to 3600 m; and in the Tien Shan (region of Khan-Tengri) up to 1900 m. Glaciers sheets of the Scandinavian type, covering the "Syrts" of the Tien Shan and the Nan Shan were widespread. Large valley glaciers descended in spots to the bottom of the depressions between the mountains and to the sloping plains of the foothills, where they formed glaciers situated at the foot of the mountains. Yet ancient glaciation was less gigantic than imagined by some of the explorers (Sinitsyn, 1959). Field observations and aerial photography show that many of the longitudinal inter-mountain valleys of the Nan Shan were filled with glaciers only partly, in their most elevated central parts (such is the case with the valleys of the rivers Tatung-ho, Peitaho and Suleho which were filled with ice in their upper reaches situated close to each other, but had only scattered small foothill glaciers near the mouths of the lateral troughs in the peripheral parts). For instance, the ancient glaciers descending from the Lenglungling range along the valley of the Ganshiga river, came out into the bottom of the wide transversal Peishui-ho valley, ending in a wide and short tongue at the altitude of 3050 m, at a distance of 10 km from the ends of the modern glaciers situated in the upper reaches of the river Ganshiga. Hypsometrically it was situated 850 m lower than these glaciers. On the terminal moraine of this ancient glacier a thick soil cover was formed, with a luxuriant steep vegetation. There are indirect indications to the effect that the retreat of the glaciers began about 5000 years ago. On the northern slope of the Tolai-shan range, ancient glaciers descended to the 2900 m mark in the basin of the Peitah river. They have left behind stone moraines lying on the intact surface of loose-like loam, on top of the pro-alluvial deposits. Present day suspended glaciers on the slopes of the ancient circular mountain wall are situated at a distance of 5-6 km from this spot, at an altitude of 4200 m.

In the Tien Shan, the ancient glaciers descended still lower. For instance the Muzart (on the eastern side of the Khan-Tengri massif) was an extremely complex huge valley glaciers which had a length of almost 100 km. It came out into the valley to an elevation of 1900 m, forming a wide (8×8 km) tongue. The glaciers of the northern slope of the Kurlyktag range also reached the foothills at an altitude of 2100 m.

The materials collected permit us to speak at most of two strictly different stages of ancient glaciation.

The traces of the reduction of present-day glaciation are numerous and testify to several (3-5 or more) stages or retreat replaced at times by advance movements.

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GLACIER-MAPPING ACTIVITIES IN THE UNITED STATES

James B. CASE

Broadview Research Corp. Washington, D. C.

SUMMARY

Precise, large-scale maps, prepared periodically, have proved very useful in determining changes in glaciers such as advance or recession, volume, rise or fall of the firn line, and so on.

A number of European countries have long had glacier-mapping programs. However, until the advent of the International Geophysical Year, such mapping in the United States was very limited. In 1957 a comprehensive two-year program of glacier mapping was begun by the American Geographical Society for the IGY program. The writer carried out ground surveys on seven glaciers in Alaska during the summers of 1957 and 1958, and, following the surveys, plotted maps from aerial photography taken in 1957.

The program has been continued through 1960 at The Ohio State University with the support of a grant from the National Science Foundation. Field work has been completed on six glaciers in the western United States and on two glaciers in Alaska. Maps of the glaciers were plotted from aerial photography taken in September of 1959. In the future, aerial photography will be taken at regular intervals so that the glaciers can be remapped and their changes determined.

In carrying out these activities, the writer has evolved a set of specifications for glacier mapping.

1. INTRODUCTION

At the Eleventh General Assembly of the International Union of Geodesy and Geophysics, held in Toronto in 1957, Dr. Richard Finsterwalder presented a paper on the "Scope, State and Development of Precise Glacier Surveys on the Earth". In this paper he outlined the reasons for which glacier maps are prepared and the methods used in their preparation; then he listed many of the glaciers throughout the world which had already been mapped, along with appropriate references. Until the advent of the International Geophysical Year, glacier-mapping activities in the United States were very limited and, as a consequence, they received little notice in Dr. Finsterwalder's paper. It will be the purpose of my paper, then, to discuss the historical background of glacier surveys on the North American continent, particularly in the United States, and to bring up to date the information on recent and current glacier-mapping projects in that area of the world.

2. HISTORICAL BACKGROUND

Glacier mapping in North America, particularly in Canada, apparently began with the work of the Alaska-Canada Boundary Commission. Otto Klotz, while working on the boundary survey, mapped the tongue of the Baird Glacier in southeastern Alaska in 1894 using photogrammetric methods⁽¹⁾. He also did some comparative studies of the glaciers of Lituya Bay, Alaska using a map dating from 1786⁽²⁾. Around the turn of the century there was a considerable amount of activity in the Canadian Rockies and Selkirks. William Sherzer mapped the Victoria, Wenkchemna, Yoho,

⁽¹⁾ Otto J. KLOTZ, «Experimental Application of the Phototopographical Method of Surveying to the Baird Glacier, Alaska,» *Journal of Geology*, Vol. 3 July-August, 1895), pp. 512-518.

⁽²⁾ Otto J. KLOTZ, «Notes on Glaciers of South-Eastern Alaska and Adjoining Territory,» *Geographical Journal*, Vol. 14, No. 5 (November, 1899), pp. 523-534.

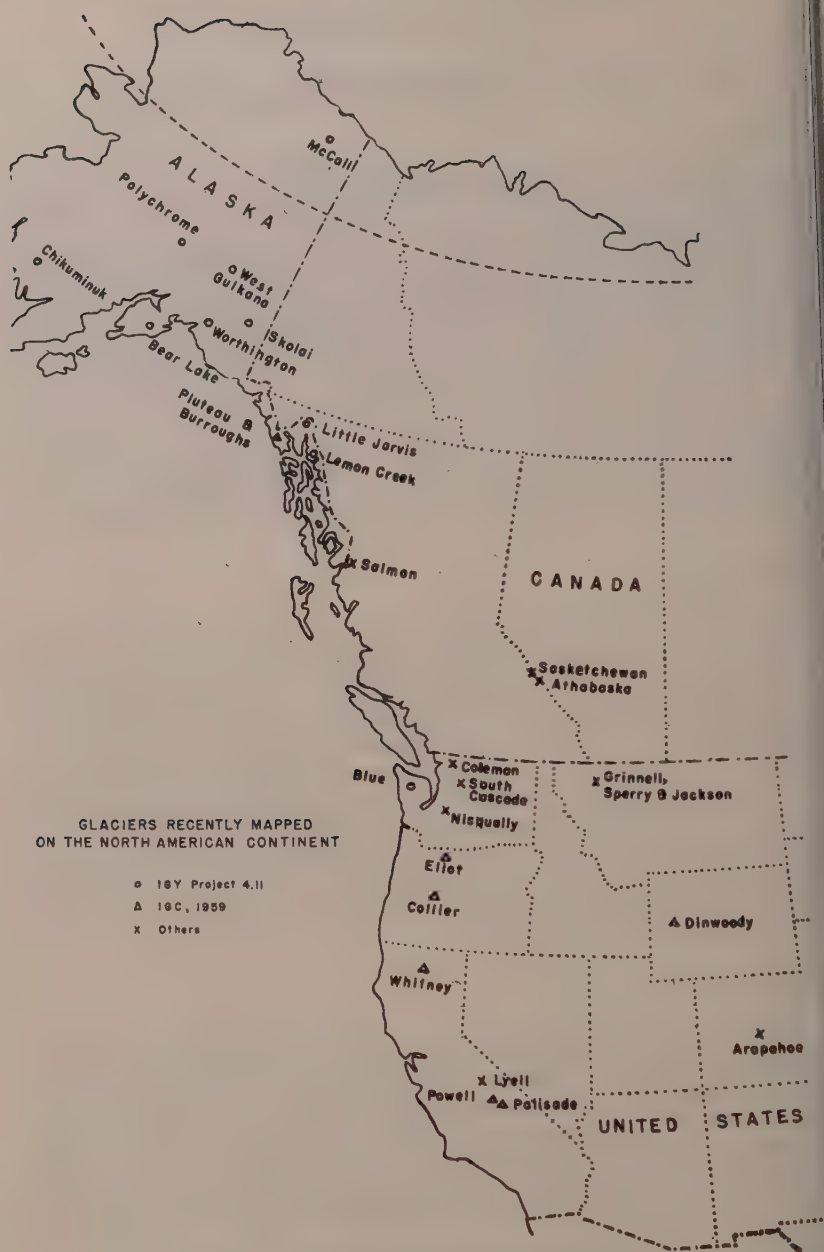


Fig. 1

Illicilliwaet, and Asulkan glaciers⁽³⁾. Work was also done on the Illicilliwaet Glacier by George Vaux and on the Yoho Glacier by A.O. Wheeler, while Howard Palmer mapped the Sir Sanford Glacier⁽⁴⁾.

Glacier mapping received much impetus in Alaska in the notable work of Ralph Tarr and Lawrence Martin on the National Geographic Society expeditions of 1909, 1910, and 1911⁽⁵⁾. These expeditions covered most of the large glaciers of northern and southeastern Alaska. This work was continued in 1931 by Wentworth and Ray and has since been expanded by the efforts of William O. Field of the American Geographical Society⁽⁶⁾. It must be noted, however, that almost all of this mapping, both in Alaska and in Canada, was of a reconnaissance nature carried out at relatively small scales. However, these maps have proved quite adequate for recording the generally rapid recessions of the glaciers in those areas.

RECENT GLACIER-MAPPING ACTIVITIES

The first serious glacier-mapping program in the United States was concentrated on the Nisqually Glacier on Mount Rainier in the state of Washington⁽⁷⁾. The mapping was begun in 1931 by the U.S. Geological Survey, in cooperation with the National Park Service, and remapping has continued every five years up to the present. The maps of 1931, 1936, 1941 and 1946, compiled from planetable surveys, and the maps of 1951 and 1956, from aerial photography, are now being published by the Geological Survey⁽⁸⁾. Concurrently, from terrestrial photography taken in 1952 and 1956, Walther Hofmann of Munich has prepared maps of the Nisqually⁽⁹⁾.

A similar, though not quite so ambitious, program has also been undertaken on several glaciers in Glacier National Park, Montana. In 1937, G.R. Gibson and L. Dyson prepared a planetable map of the Grinnell Glacier, in 1938 the Sperry Glacier, and in 1939 the Jackson. Maps of the Grinnell and Sperry glaciers were prepared again in 1946 by planetable and all three glaciers were mapped from aerial

⁽³⁾ William H. SHERZER, «Glaciers of the Canadian Rockies and Selkirks», *Smithsonian Contributions to Knowledge*, Vol. 34 (1907).

⁽⁴⁾ George VAUX and William S. VAUX, Jr., «The Great Glacier of the Illicilliwaet», *Appalachia*, Vol. 9, No. 2 (March, 1900), pp. 156-165; A.O. Wheeler, «Motion of the Yoho Glacier», *Canadian Alpine Journal*, Vol. 1, No. 2 (1908), pp. 271-275; Howard Palmer, «Observations on the Sir Sanford Glacier, 1911», *Geographical Journal*, Vol. 39, No. 5 (May, 1912), pp. 446-453.

⁽⁵⁾ Ralph S. TARR and Lawrence MARTIN, *Alaskan Glacier Studies*, The National Geographic Society, Washington, 1914; see also Lawrence Martin, «Gletscheruntersuchungen längs der Küste von Alaska», *Petermanns Mitteilungen*, Vol. 58 (1912), pp. 78-81.

⁽⁶⁾ C.K. WENTWORTH and L.L. RAY, «Studies of Certain Alaskan Glaciers in 1931», *Bulletin, Geological Society of America*, Vol. 47 (June 30, 1936), pp. 879-933; William O. Field, «Observations on Alaska Coastal Glaciers in 1935», *Geographical Review*, Vol. 27, No. 1 (January, 1937), pp. 63-81; William O. Field, «Glacier Recession in Muir Inlet, Glacier Bay, Alaska», *Geographical Review*, Vol. 37, No. 3 (July, 1947), pp. 369-399.

⁽⁷⁾ V.R. BENDER and A.L. HAINES, «Forty-two Years of Recession of the Nisqually Glacier on Mount Rainier», *Erdkunde*, Vol. 9, No. 4 (December, 1955), pp. 275-281.

⁽⁸⁾ The planetable maps are at a scale of 1:9,600, the others 1:12,000.

⁽⁹⁾ Walther HOFMANN, «Photogrammetric Glacier Measurements on the Volcanic Peaks of Washington», *The Mountaineer*, Vol. 46, No. 13 (December 15, 1953), pp. 7-16; Walther Hofmann, «Der Vorstoss des Nisqually-Gletschers am Mt. Rainier, USA, von 1952 bis 1956», *Zeitschrift für Gletscherkunde und Glazialgeologie*, Vol. 4, No. 1-2 (1958), pp. 47-60. A scale of 1:25,000 and contour interval of 20 meters was used.

photography in 1950⁽¹⁰⁾. The most recent mapping was completed by Arthur Johnson of the U.S. Geological Survey in 1956 on the Sperry and 1957 on the Grinnell with terrestrial photography⁽¹¹⁾.

Another glacier which has received much attention in recent years is the Coleman on Mt. Baker in the state of Washington. It has been mapped from terrestrial photography every year since 1954 by A.E. Harrison of the University of Washington⁽¹²⁾. Similarly, Oliver Kehrlein began planetable surveys of the Powell Glacier in the Sierra Nevada of California in 1953, and these surveys were repeated every year through 1957⁽¹³⁾.

A number of other glaciers have been mapped though not as a part of any regular program. For example, the Palisade Glacier in the Sierra Nevada was mapped in 1946; the Lyell Glacier, also in the Sierra Nevada, was surveyed in 1957 and 1959; a map of the South Cascade Glacier was prepared in 1955 from aerial photography; and a planetable map of the Arapahoe Glacier in Colorado is presently being prepared⁽¹⁴⁾.

Recent glacier mapping in Canada has been very limited. Mark Meier mapped parts of the Saskatchewan Glacier in 1948 and 1952⁽¹⁵⁾; the National Research Council of Canada has recently published a map of the Salmon Glacier, prepared from aerial photography taken in 1949 and 1957; and the Department of Northern Affairs and National Resources is presently compiling a map of the Athabaska Glacier from aerial photography taken in 1959. From Mexico, José L. Lorenzo reports that medium-scale maps of the glacier-covered areas of Citlaltepetl, Popocatepetl, and Iztaccihuatl have been prepared⁽¹⁶⁾.

4. GLACIER-MAPPING PROGRAMS OF THE IGY AND THE IGC, 1959

In 1948 the American Geographical Society initiated work on the Juneau Icefield Research Project under Office of Naval Research sponsorship. In 1955 a control survey and some terrestrial photography was completed on the Lemon Creek Glacier,

⁽¹⁰⁾ James L. DYSON, «Shrinkage of Sperry and Grinnell Glaciers, Glacier National Park, Montana», *Geographical Review*, Vol. 38, No. 1 (January, 1948), pp. 95-103; Arthur Johnson, «Investigations on Grinnell and Sperry Glaciers, Glacier National Park, Montana», *Extrait des Comptes Rendus et Rapports—Assemblée Générale de Toronto 1957*, Vol. 4, pp. 525-534. Scales of 1:2,400 and 1:4,800 and a contour interval of 20 feet were used.

⁽¹¹⁾ Arthur JOHNSON, *Glacier Observations—Glacier National Park, Montana—1957 and 1958*, U.S. Geological Survey, Open File, December, 1958.

⁽¹²⁾ Kermit B. BENGTON, «Activity of the Coleman Glacier, Mt. Baker, Washington, USA, 1949-55», *Journal of Glaciology*, Vol. 2, No. 20 (October, 1956), pp. 708-713; J.E. Colcord, «The TAF Phototheodolite and Its Use in Glacier Surveys», *Photogrammetric Engineering*, Vol. 23, No. 3 (June, 1957), pp. 552-557. A scale of 1:5,000 and contour interval of 50 feet are being used.

⁽¹³⁾ This project was begun by the Glacier Study Subcommittee of the American Geophysical Union and the Sierra Club and was continued by the U.S. Geological Survey. Scales of 1:1,200 and 1:2,400 with 10- and 20-foot contour intervals were used.

⁽¹⁴⁾ The Palisade Glacier was mapped at a scale of 1:6,000 and contour interval of 50 feet by planetable. See Weldon F. Heald, «Palisade Glacier Survey, Sierra Nevada», *American Alpine Journal*, Vol. 6, No. 3 (1947), pp. 332-339. The Lyell Glacier, mapped at a scale of 1:2,400 with a contour interval of 20 feet, and the South Cascade Glacier, mapped at a scale of 1:5,400 with a contour interval of 20-40 feet, were both prepared by the U.S. Geological Survey. The map of the Arapahoe Glacier, at a scale of 1:1,800 and contour interval of 20 feet, is being prepared by Henry Waldrop of the University of Colorado.

⁽¹⁵⁾ MARK F. MEIER, *Mode of Flow of Saskatchewan Glacier, Alberta, Canada*, U.S. Geological Survey Professional Paper 351, U.S. Government Printing Office, Washington, 1960.

⁽¹⁶⁾ José L. LORENZO, *Los Glaciares de México*, Instituto de Geofísica, Universidad Nacional Autónoma de México, México, D.F., 1959.

a part of the Juneau Icefield, for the purpose of preparing a map of that glacier. During this same period a glacier-mapping program for the International Geophysical Year was conceived. Administered by the American Geographical Society, it was to include a number of glaciers distributed throughout the various climatological regions of Alaska. The Lemon Creek Glacier was one of these. Active planning for this program, known as IGY Project 4.11, began in 1956. I was asked to prepare the maps. This included planning the aerial photography, carrying out field surveys and terrestrial photography and, finally, plotting the maps⁽¹⁷⁾.

The glaciers to be mapped were selected on the basis of accessibility, size (less than 8 kilometers in length), and simplicity of shape. A tentative selection was made by William Field of the American Geographical Society from available maps and aerial photographs as well as from his own intimate knowledge of the Alaskan glaciers. Austin Post, field-party leader during the first season's work, made some changes in the glaciers chosen. The locations of the glaciers finally decided upon are shown on the accompanying map. The plans called for a field party of three men to spend the summers of 1957 and 1958 in Alaska completing the ground surveys and terrestrial photography necessary for preparing topographic maps of the glaciers at a scale of 1:5,000 with a 5-meter contour interval. At the same time, arrangements were made for the U.S. Navy to obtain aerial photography of all the glaciers during the summer of 1957.

The first summer's field party consisted of Austin Post as leader, myself as photogrammetrist, and Richard Long as assistant. We were able to complete the work on three glaciers that summer: the Worthington, West Gulkana, and Polychrome. All were within easy walking distance of roads so that we were able to use a light truck for transportation. We would normally pitch camp part-way up the glacier where we would live, barring bad weather, until the survey was completed. The survey consisted in measuring a base line using a 2-meter subtense bar, and then expanding by the usual triangulation procedures to one or two quadrilaterals straddling the glacier. In addition, a number of cairns and prominent peaks would be intersected. Finally, terrestrial photographs were taken from suitably positioned base lines. A Wild phototheodolite was used both for triangulation and photography. An aneroid altimeter was used to obtain an approximate elevation datum, but no attempt was made to establish a position or azimuth datum.

The second summer saw some changes in personnel and procedures. I took over as leader, Richard Long was again assistant, and a new assistant, Wayne Merry, joined the group. We completed surveys on four glaciers: the Bear Lake (formerly the Salmon Creek), Chikuminuk (formerly Kilbuck), Skolai, and Little Jarvis. This time the glaciers were more difficult of access. A light plane equipped with floats landed us on a lake near the Chikuminuk Glacier after a flight of about 100 kilometers. Two weeks later the plane returned for us. Much the same situation existed with the Skolai and Little Jarvis glaciers except that there were small air strips near the glaciers where the planes could land. In every case, we took advantage of air drops for getting supplies into our camp areas. I decided not to take terrestrial photography the second summer because of unsatisfactory results obtained the previous summer and because it was necessary to travel as lightly as possible. Therefore, a one-second theodolite was used for triangulation and a 500-foot tape for base line measurements.

The Wild A-7 Autograph at the Division of Geodetic Sciences of The Ohio State University was made available to me for plotting the glacier maps. Standard photogrammetric procedures were used and no difficulties were encountered in the plotting. Scales ranged from 1:5,000 to 1:10,000. A contour interval of 5 meters was used on the glacier and moraines, and a 25-meter interval was used elsewhere. A map

(17) James B. CASE, «Mapping of Glaciers in Alaska», *Photogrammetric Engineering*, Vol. 24, No. 5 (December, 1958), pp. 815-821.

of the Skolai Glacier was not prepared because the Navy had not been able to obtain suitable aerial photography. Otherwise, maps were plotted for all of the glaciers in the program. The terrestrial photography taken on the Lemon Creek Glacier in 1955 proved unsatisfactory for mapping so the plotting of that map was also done from aerial photography taken by the U.S. Navy.

I should mention here two additional glaciers that were mapped as a part of the American Geographical Society's program although the field surveys and aerial photography were done by other organizations. These were the McCall Glacier in the Brooks Range of Alaska and the Blue Glacier in the Olympic Mountains in the state of Washington.

All of these maps are presently being published by the American Geographical Society ⁽¹⁸⁾.

Early in 1959, at the suggestion of William Field, I prepared a program for mapping a number of glaciers in the western United States. The proposal, submitted directly to the National Science Foundation, included an initial program for carrying out field surveys, obtaining aerial photography, and plotting maps for the following glaciers: Palisade and Powell in the Sierra Nevada of California; Whitney on Mt. Shasta in California; Collier on the Three Sisters and Eliot on Mt. Hood, both in Oregon; Dinwoody in the Wind River Range of Wyoming; and Plateau and Burroughs in Glacier Bay National Park, Alaska. The selection of glaciers was made on the basis of areas where work had previously been done and on the basis of geographical distribution. Suggestions from Oliver Kehrlein of the California Academy of Sciences and Mark Meier of the U.S. Geological Survey helped in the selection of particular glaciers. The two glaciers in Alaska were to be mapped in support of a project being carried out by Richard Goldthwait of the Department of Geology of The Ohio State University.

The field party, consisting of myself and Richard Long, began the survey work at the beginning of July, 1959 and we were finished by the end of August. We traveled between glaciers by automobile (except in Alaska) and were able to come within walking distance of all of them. We did find it expedient to hire pack animals for several of the glaciers. In general, the survey methods were identical to those we had used in Alaska the previous summer. We met no difficulties and found we could complete the work on a glacier in about two days. However, we were not so fortunate with respect to the aerial photography. Bad weather and an early winter snowfall prevented the acquisition of photography in Alaska and on the Palisade and Dinwoody glaciers. The rest of the photography was successfully flown and I was able to complete the plotting of the maps of the four remaining glaciers ⁽¹⁹⁾.

5. FUTURE GLACIER MAPPING

When I submitted the proposal for this work to the National Science Foundation, I also included some suggestions for a long-range glacier-mapping program. Obviously the maps already prepared under IGY Project 4.11 in Alaska and the IGC program in the western United States are practically useless for determining such things as volume change, advance or recession, rise or fall of the firn line, and so on, unless follow-up maps are prepared with which comparisons may be made. Therefore, any future glacier-mapping program should include the flying of aerial photography and plotting of maps at regular time intervals, say every five years, for all of the glaciers

⁽¹⁸⁾ AMERICAN GEOGRAPHICAL SOCIETY, *Nine Glacier Maps*, Special Publication no. 34, American Geographical Society, New York, 1960.

⁽¹⁹⁾ JAMES B. CASE, *Glacier Mapping in the Western United States*, Report 943, The Ohio State University Research Foundation, Columbus, January, 1960.

which have already been mapped. The decision concerning the time for this remapping, I believe, to be worked out at the Helsinki Assembly. I also suggested that, in the mean-time, steps be taken to prepare maps of these glaciers from existing aerial photography. In many instances aerial photography suitable for mapping has been flown over these areas as many as two or three different times in past years. Maps drawn from this photography would, when compared with the maps already prepared, give a very good indication of changes that have occurred.

CONCLUSION

It has been my intention in this paper to describe some of the early history of glacier mapping in North America and to bring up to date information on recent work. I have also gone into some detail on the two projects with which I have been associated since they represent a rather comprehensive beginning to a long-term program which envisions the regular remapping of those glaciers.

RAPPORT SUR LES VARIATIONS DE LONGUEURS DES GLACIERS EUROPEENS, EN 1956/57, 1957/58, 1958/59

présenté à L'ASSEMBLÉE DE L'UNION GÉODÉSIQUE ET GÉOPHYSIQUE À HELSINKI
par L'ASSOCIATION INTERNATIONALE D'HYDROLOGIE

Ce Rapport est le huitième du même genre rédigé par le soussigné depuis 1914 d'abord pour la Commission Internationale des Glaciers, (CIG), puis, dès la fusion (Oslo 1948) de cette commission avec celle de la Neige en un seul organisme : la Commission Internationale de la Neige et de la Glace (CING), sous l'égide de J.E. Church et de P.L. Mercanton, présidents honoraires de la C.I.N.G.

Le présent rapport triennal est lui aussi le cinquième rédigé pour la CING par le soussigné, qui l'a établi dans sa forme usuelle, seule actuellement praticable jusqu'à ce qu'un accord général soit réalisé pour le contrôle des glaciers du monde entier.

Voici les noms des correspondants nationaux du contrôle des glaciers européens publiés ici. Comme d'usage, nos collaborateurs gardent la responsabilité des données fournies. Qu'ils veuillent bien trouver ici les remerciements du rapporteur général comme de la CING toute entière. Le rédacteur exprime à cette occasion le regret que force renseignements intéressants sur les changements de volume ou d'aire des glaciers de certains pays, la France notamment, n'aient pu être encadrés dans la présentation du Rapport.

Les variations frontales des glaciers sont indiquées ici en mètres, au demi-mètres près. La décrue est indiquée par le signe —, le stationnement par un 0, la crue par un +; un X remplace le chiffre défaillant éventuel.

Pour la CING, le rapporteur : P.L. Mercanton.

Le sous-signé remercie tout particulièrement la Commission Helvétique des Glaciers, pour son appui moral et matériel constant, ainsi que M^{me} Verneuil de Marval pour son aide dévouée à la mise en forme d'impression de ce rapport comme d'ailleurs des précédents.

Pour la C.I.N.G.
Paul L. Mercanton
Président honoraire

VARIATIONS DE LONGUEUR DES GLACIERS DES ALPES FRANÇAISES, EN MÈTRES

<i>Haute Savoie</i>	1956/59 (3 ans)	
La Tour	+ 35	»
Argentière	— 107	»
Mer de Glace	— 30	»
Les Bossons	+ 95	»
Taconnaz	+ 12	»
Bionnassay	— 35	»
Tré la Tête	— 12	»
	1955/59 (4 ans)	
Glacier de Gébroulaz	— 55	
<i>Isère</i>	1956/57	1957/59 (2 ans)
Glacier de la Pilatte	— 11	— 74
	1951/57 (6 ans)	
Glacier de la Salle	— 100	

<i>Hautes Alpes</i>	1956/58 (2 ans)	1958/59
Glacier Blanc	— 19 »	— 23
Glacier Noir	— 2 »	+ 32

Hautes Pyrénées

Des trois glaciers du Vignemale (Aussoue, Oulettes de Gaube, Petit Vignemale), observés de 1956 à 58 (2 ans), 2 étaient en décrue, 1 stationnaire.

Observations recueillies par M^llrs. Bouverot, Anchierri, Huin, Rogie, Sannac, Chimits.

VARIATIONS DE LONGUEUR DES GLACIERS DES ALPES SUISSES, EN MÈTRES

<i>Bassin du Rhône</i>	1956/57	1957/58	1958/59
Glaciers :			
Glacier Rhône	—	— 2 (2 ans)	— 8
Glacier Mutt	— 14	— 10	—
Glacier Mesch	— 2,5	— 5	— 1,5
Glacier Grosser Aletsch	— 24	— 20,5	— 10
Glacier Fäng	— 5	— 12	— 14
Glacier Kaltwasser	— 17,5	— 35,5	— 17,5
Glacier Allalin	+ 17,5	+ 2	+ 2
Glacier Illiboden	— 10	— 0,5	— 0,5
Glacier Schwarzberg	— 7,5	— 10	— 8
Glacier Fental	— 5,5	— 2,5	— 8
Glacier Fessjen	— 2	— 12	— 7
Glacier Fée	—	—408 (5 ans)	+ 9
Glacier Fied	—	— 9 (2 ans)	— 24
Glacier Forner	— 36	— 20,5	— 56
Glacier Mutt	— 24 (2 ans)	— 36	— 31,5
Glacier Fendelen	— 24 »	—478	— 41
Glacier Furtmann (Ouest)	—11,5	— 66	— 9,5
Glacier Furtmann (Est ou Brunegg)	— 8	— 5	— 7
Glacier Frenal	— 8	— 10,5	— 13,5
Glacier Foming	— 8,5	— 10,5	— 16
Glacier Fella Tola	+ 21,5 (2 ans)	+ 4	— 31,5
Glacier Foiry	— 6	— 14	— 6,5
Glacier Ferpècle	— 34	— 42	—
Glacier Font Miné	— 6,5	— 54,5	—
Glacier Fals d'Arolla (Arolla)	— 23,5	— 10,5	— 22
Glacier Faldjore-Nouve	— 9	— 12	— 5
Glacier Fheilou	— 18	— 2,5	— 6,5
Glacier Fend Darrey	+ 27	— 41	— 70
Glacier Grand Désert	— 7	— 7	— 26
Glacier Font Fort	— 2,5	— 6	— 10,5
Glacier Femma	0	— 22 (2 ans)	— 15
Glacier Font Durand (Val de Bagnes)	— 30 (2 ans)	— 53	—
Glacier Fendey	— 16	— 14	— 13
Glacier Falsorey	— 22	— 15	— 31
Glacier Fendet	— 14	— 5	— 20
Glacier Furbassière	— 15,5	— 46	— 12
Glacier Fétro	— 0,5	—	— 0,5 (2 ans)

VARIATIONS DE LONGUEUR DES GLACIERS DES ALPES SUISSES (Suite)

	1956/57	1957/58	1958/59
Glaciers :			
Saleina	— 9	— 11	— 46
Trient	— 37	— 43	—
Tsanfleuron	— 6	— 39,5	— 81,5
Plan Névé (Grand)	— 11 (2 ans)	—	— 3 (2 ans)
Martinets	— 8 »	+ 1	— 4
Prapio	— 12	+ 1,5	— 2,5
Sex Rouge.	— 3	— 1	— 3,5
Paneyrosse.	— 2 (2 ans)	+ 1,5	—
<i>Bassin de l'Aar</i>			
Oberaar*	— 44	— 37	— 60
Unteraar*	— 5	— 2	— 6,5
Oberer Grindelwald	— ×	— ×	— ×
Unterer Grindelwald	— 14	— 16	— 38
Stein	— 8	— 7	— 22
Blümlisalp	— 12 (2 ans)	—	—
Schwarz	— 1	— 2	—
Gamchi	— 3	— 13	—
Rätzli	— 43	— 47	— 17
Wildhorn	+ 2	— 1	—
Gauli	—	—	—
<i>Bassin de la Reuss</i>			
Griess (bei Unterschächen)	—	— 5 (3 ans)	+ 2
Wallenbühl (bei Voralp)	— 21	— 27,5	—
Chelen	— 22,5	— 13,5	— 23
Rotfirn	— 7,5	— 26	+ 12,5
Hüfi	— 25	— 37	—
Brünni	—	— 8 (2 ans)	—
Damma	— 22,5	+ 12	—
St. Anna.	— 11	— 8	+ 3,5
Tiefen.	— 11,5	— 5,5	— 16,5
Firnälpli Ost (Grassengletscher)	+ 11,5	— 31	—
Griess (Griessengletscher)	+ 9	— 29,5	—
<i>Bassin de la Linth</i>			
Sulz.	+ 2,5 (2 ans)	— 7	—
Biferten	— 90 (8 ans)	— 2,5	— 4
Glärnisch	—	—	—
<i>Bassin du Rhin</i>			
Punteglias	— 1,5	— 9	— 8,5
Vorab.	—	—	— 21,5 (3 ans)
Lavaz	— 28,5 (2 ans)	— 25	— 3

*En décrue certaine, mais accentuée par l'ennoyage estival du front

VARIATIONS DE LONGUEUR DES GLACIERS DES ALPES SUISSES (Suite)

	1956/57	1957/58	1958/59
Porchabella	— 8,5	— 22	—
Verstankla	+ 43 (2 ans)	— 11	— 20
Silvretta	— 8	— 9	— 18
Lenta	— 56	— 7,5	— 8
Sardona	0	— 12	— 7
Paradies	— 33,5	— 16	— 31
Suretta	— 51	— 16	— 16
Pizol	— 0,5	— 38	— 37

Bassin de l'Inn

Morteratsch	— 31	— 45,5	— 39
Tschierva	— 117 (2 ans)	— 40	— 18
Tiatscha	+ 9,5	— 14	+ 14
Sesvena	— 4	— 8,5	— 5,5
Calderas	— 17,5 (2 ans)	— 8	— 17
Lischana	+ 14	— 30,5	— 10
Roség	— 77,5 (2 ans)	— 32,5	— 24,5

Bassin de l'Adda

Forno	— 31	— 28	—
Palù	— 23,5	— 20,5	— 20,5
Paradiso	0	— 15	+ 1
Cambrena	— 5	— 2,5	— 3

Bassin du Tessin. . . .

Rosshoden	— 2,5 (2 ans)	— 2	— 11
Bresciana	— 18,5	— 22	— 22
Basodino	— 18	— 7,5	— 33

VARIATIONS DE LONGUEUR DES GLACIERS DES ALPES AUTRICHIENNES, EN MÈTRES

Gletscher, Ferner, Kees	1956/57	1957/58	1958/59
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Hochkönig

«Übergossene Alm» . . .	— 0,5	— 6	— 5
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Dachstein («Gletscher»

Gr. Gosau	— 4,5	— 14,5	— 1,5
Hallstätter	— 1,5	— 3	— 15
Schladminger	— 0	— 3	— 3

Silvrettagruppe («Ferner»)

Litzner NE	— 5	— 9	— 14
Klostertaler	— 2	— 2	— 14
Ochsentaler	—	— 20	— 20

VARIATIONS DE LONGUEUR DES GLACIERS DES ALPES AUTRICHIENNES, (Suite)

Gletscher, Ferner, Kees	1956/57	1957/58	1958/59
Fermunt W	— 20,5	— 14,5	} — 3
Fermunt E.	— 6	—	
Bieltaler	— 18	— 11,5	
Jamtaler	— 5,5	— 11,5	— 6
Larein.	—	—	—

Otztaler Alpen («Ferner»)

Weissee	— 11	— 14,5	— 19,5
Gepatsch	— 41,5	— 18	— 28,5
Hinterer Olgruben	—	— 2	— 2
Sexegeten	— 7	— 13,5	— 15,5
Taschach	— 18,5	— 31	— 27,5
Mittelberg	— 13	— 8	— 4,5
Karles	— 8	— 3	— 3,5
Hochjoch	— 10	— 26	— 20
Hintereis	— 12	— 23	— 100
Guslar	— 8,5	— 31,5	— 29
Vernagt	— 14,5	— 23	— 22
Mitterkar	0	—	—
Rofenkar	— 1	—	—
Taufkar	—	—	—
Niederjoch	— 15,5	— 24	— 19,5
Marzell	0	— 21,5	— 16
Schalf	— 25	— 15	— 5
Diem	— 23,5	— 19	— 26,5
Spiegel	— 7,5	— 7,5	— 26,5
Gurgler	— 16	— 16	— 30
Langtaler	— 10	— 10	— 16
Rotmoos	— 11	— 16	— 16
Gaisberg	— 16,5	— 16	— 16

Stubaiier Alpen («Ferner»)

	1956/58 (2 ans)		
Sulztaler	— 82,5	»	—
Bockkogel	— 94	»	—
Schwarzenberg	— 30	»	—
Bachfallen	— 5	»	—
Längentaler	— 0	»	—
Lisenser	— 16,5	»	—
Alpeiner	— 19	»	—
Berglas	— 30	»	—
Hochmoos	— 9	»	— 9
Daunkogel	—		— 22,5 (3a)
Schaufel	+ 2	»	— 0
Fernau	— 18,5	»	—
Sulzenau	— 23	»	— 25,5

VARIATIONS DE LONGUEUR DES GLACIERS DES ALPES AUTRICHIENNES, (Suite)

Gletscher, Ferner, Kees	1956/57	1957/58	1958/59
Grünau	—	1 »	— 1
Grübl W	—	—	—
Grübl E	—	8 »	— 3
Simminger	—	18 »	—
<i>Zillertaler Alpen (Kees)</i>		1957/58	
Waxegg	— 4	— 8	—
Horn	— 23,5	— 29,5	—
Schwarzenstein	— 5	— 4,5	—
<i>Glocknergruppe («Kees»)</i>			
Karlinger	—	— 39 (2 ans)	—
Bärenkopf	—	—	—
Klockerin	—	—	—
Pasterzen	— 9,5	— 10	— 17,5
Wasserfall	— 3	— 32	—
Freiwand	—	—	—
Pfandscharten	—	—	—
<i>Ankogelgruppe («Kees»)</i>			
Grosselend	— 8,5	— 15	— 13,5
Kleinlend	— 8,5	—	— 1,5
Kälberspitz	— 11,9	—	— 12,5
Tripp W.	—	— 3	— 3
Hochalm	— 1	— 4,5	— 1,5
Winkel	— 11	— 3,5	+ 2,5

Observations recueillies par MMrs. Wannenmacher, Schueller, Prutzer, Mutschlechner, Schatz, Held, Heuberger, Lässer, Paschinger, Pacher.

VARIATIONS DE LONGUEUR DES GLACIERS ISLANDAIS, EN MÈTRES

	1956/57	1957/58	1958/59
<i>Drangajökull</i>			
Kaldalonsjökull	— 18	— 48	— 48
Leirufjardarjökull	— 30	—	—
Reykjafjardarjökull	— 26	— 15	— 20
<i>Snoesfellsjökull</i>			
Hyrningsjökull	— 6	—	—
Jökulhals	— 48	—	—

VARIATIONS DE LONGUEUR DES GLACIERS ISLANDAIS, (*Suite*)

	1956/57	1957/58	1958/59
<i>Eyjafjalla-Myrdalsjökull</i>			
Gígjökull	—	— 181 (4 ans)	—
Solheimajökull, V-spordur (W-snout)	—	— 37 (2 ans)	—
» » » A-spurdur (E-snout)	—	— 30 »	—
» » » Jökulhöfud	—	— 5 »	—
<i>Vatnajökull</i>			
Skeidararjökull vestan til (W-side)	— 63	— 60	— 130
Skeidararjökull austan til (E-side)	— 24	— 28	— 12
Morsarjökull	— 14	— 39	— 27
Skaftafellsjökull	+ 6	+ 2	— 35
Svinafellsjökull nordan til (N-side)	+ 3	— 17	— 13
Svinafellsjökull sunnan til (S-side)	— 20	+ 1	— 18
Virkisjökull	— 15 (2 ans)	— 8	— 8,5
Kviarjökull	— 6	— 20	— 35
Hrutarjökull	— 11	— 14	— 12
Fjallsjökull	+ 5	0	— 8
Breidamerkurjökull vestan ar (W of Jökulsa)	— 23	— 30	— 54
Breidamerkurjökull austan ar (E of Jökulsa)	— 75	— 58	— 63
Fellsarjökull	—	— 60 (2 ans)	—
Brokarjökull	— 15	— 10	— 20
Birnujökull	— 1	— 3	— 2
<i>Heinabergsjöklar</i>			
A Eyvindstungnakolli	— 5	— 5	— 17
Sydri jökull (Skalafellsjökull)	— 29	— 13	— 15
Nýrdri jökull (Heinabergs- jökull)	— 37	— 13	+ 30
Flaajökull	— 15	— 37	+ 22
Hoffellsjökull, eystri tunga. (E-snout)	— 6	+ 3	+ 5
Hoffellsjökull, vestri tunga (W-snout)	— 53	— 49	— 90
Tungnarjökull (Jökulhei- mum)	— 40	— 40	— 50
<i>Hofsjökull</i>			
Nauthagajökull	+ 86	— 13	— 77
Mulajökull	— 2	— 72	— 29
A Lambahrauni (NV-horn)	—	— 30	—

Langjökull

Fulakvisl (Pjofad)	—	— 70 (2 ans)	— 60
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Hrutafell

Midjökull	—	— 77 (7 ans)	— 3
Vesturjökull	—	— 45 » »	— 21
Nordvesturjökull	—	— 70 » »	0

Kerlingarfjöll

Lodmundarjökull (innri)	—	— 2 (2 ans)	+ 3
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Nordurlandssjöklar . .

Gljufurarjökull	— 16	— 6	— 21
Tungnahryggsjökull, austan ar (E-side)	—	— 107 (6 ans)	—
Tungnahryggsjökull, vestan ar (W-side)	—	— 71 » »	—

Observations recueillies par Mr. Jon Eythorsson

VARIATIONS DE LONGUEUR DES GLACIERS SUÉDOIS, EN MÈTRES

	1956/57	1957/58	1958/59
Stor glaciären	— 10	— 7	— 4
Isfall	— 14	— 12	— 8
Kaskasat joko SO	— 12	— 10	— 6

VARIATIONS DES GLACIERS NORVÉGIENS, EN MÈTRES

<i>Jostedalsbreen</i>	1956/57	1957/58	1958/59
Bøyumbreen	— 5	+ 10	0
Store Suphellebreen	— 2	+ 5	0
Austerdalsbreen	— 47 (2 ans)	— 43	— 38
Tunsbergdalsbreen	— 6	— 5	— 15
Nigardsbreen	— 34	— 49	— 66
Fäbergstølbreen	— 36	— 60	— 46
Lodalsbreen	— 27	— 47	— 48
Stegaholtbreen	— 25	— 34	— 50
Briksdalsbreen	+ 13	+ 27	+ 14
Abrekkebreen	— 4 (2 ans)	+ 9	+ 12

Jotunheimen

Tverrabreen	— 10	— 2	— 43
Veslejuvbreen	— 2	0	— 4
Storjuvbreen	— 31	— 17	— 78
Heimre Illabre	— 14	—	— 27 (2 ans)
Nordre Illabre	— 14	— 13	— 21
Søndre Illabre	— 18	— 22	— 25

Storbreen (Leirdalen)	— 20	—	— 40 (2 ans)
Leirbreen	— 10	—	— 25 (2 ans)
Böverbreen	— 36	—	— 26 »
Slettmarkbreen	—	— 14 (2 ans)	—
Svartdalsbreen	—	— 30 »	—
Langedalsbreen	—	— 12 (2 ans)	—
Hellstugubreen	— 13	— 4	— 39
Styggedalsbreen	— 3	— 8	— 15,5
Styggebreen	— 30	—	55 (2 ans)

Möre

Trollkyrkjebreen	+ 12	— 11	— 12
Veslebreen	+ 7 (3 ans)	— 4	— 8
Finnebreen	—	—	— 17 (3 ans)
Kolasbreen	—	—	— 5 »

Folgefonn

Bondhusbreen	—	+ 14 (2 ans)	— 16
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VARIATIONS DE LONGUEUR DES GLACIERS DES ALPES ITALIENNES, EN MÈTRES

Alpi Occidentali

<i>Alpi Marittime</i>	1956/57	1957/58	1958/59
Ghiacciaio Clapier	— 1,5	— 1	— 0,5
» » Peirabroc	— 1	— 0,5	—
» » Maledia	— 1,5	—	—
» » Murajon	— 1	— 10	+ 0,5
» » Gelas	— 0,5	— 1	—

Alpi Scozie

Ghiacciaio Monviso N E.	— 2	—	—
» » Calambra Or. Le	— 6,5	—	—
» » Calambra Occ. Le	— 12	—	—

Alpi Graie

Ghiacciaio Bessanese	+ 2	— 16	—
» » Mulinet Nord.	— 11	—	—
» » Mulinet Sud	— 4	—	—
» » Sea	— 25	—	—
» » Capra	— 7	— 8	— 15
» » Carro	—	—	— 5 (3 ans)
» » Nel Or. Le	+ 5	— 3	—
» » Etrét	—	—	—
» » Monciar	— 4	— 10,5	— 18,5
» » Moncorve	— 5	— 5,5	— 10
» » Gr. Paradiso	—	—	—
» » Coletta	— 17,5	— 17	— 21,5
» » Trayo	— 2	—	—
» » Chavannes	— 5	—	— 2,5 (2 ans)
» » Arguerey	—	—	—

VARIATIONS DE LONGUEUR DES GLACIERS DES ALPES ITALIENNES (*Suite*)

Ghiacciaio	1956/57	1957-58	1958-59
» » Pré de Bar . . .	— 7	— 3	— 3
» » Triolet	— 2	— 2,5	— 2,5
» » Trebuzie	—	— 10 (2 ans)	— 1
» » Jorasse	—	—	—
» » Planpincieux . .	—	—	—
» » Rochefort . . .	—	—	—
» » Frety	— 1	— 1,5	— 2
» » Toulà	— 8	— 6	— 8
» » Entrèves	— 3	— 1	— 1,5
» » Brenva	—	— 10 (2 ans)	— 15
» » Frenay	—	— 1 (2 ans)	—
» » Brouillard . . .	—	—	—
» » Miage	—	— 5 (2 ans)	—
» » Lée Blanche . .	— 4	—	—
» » Estellette . . .	—	— 3 (2 ans)	—

Alpi Centrali

Alpi Pennine

Ghiacciaio Tra de Tran . .	— 6	— 38	— 35
» » Montabel	—	—	—
» » Cherillon	+ 4	— 2,5	— 1
» » Valtournanche . .	— 10	— 15	— 15
» » Furca	—	— 8 (2 ans)	— 12
» » Tyndall	—	—	—
» » Cervino	—	—	—
» » Grande Verra . .	— 3	— 14	— 12
» » Piccolo Verra—	— 4	— 17	— 63
» » Lys	— 12,5	— 21,5	— 6,5
» » Indren	— 7	— 6	—
» » Piode	— 6	—	—
» » Sesia	— 10	— 20	— 10
» » Belvédère	— 22	— 21	— 34
» » Nordend	—	— 5,5 (2 ans)	— 3,5

Alpi Retiche

Ghiacciaio Ferre	— 20	— 30	— 17
» » Val Loga	— 8	— 10	— 9
» » Suretta	— 15	— 17	— 13
» » Orsareigls	— 7	— 8	— 6
» » Ponciagna	—	— 9 (2 ans)	— 8
» » Ventina	— 20	— 30	— 19
» » Caspoggio	— 12	— 20	— 6
» » Fellaria	— 18	— 36	—
» » Scherschen	— 70	—	—
» » Scalino	— 17	— 20	— 34
» » Rinalpi	— 2	— 3	— 2
» » Piazzai Or. Le . .	— 1,5	—	— 15 (2 ans)

VARIATIONS DE LONGUEUR DES GLACIERS DES ALPES ITALIENNES (Suite)

	1956/57	1957/58	1958/59
Ghiacciaio Dosde' Or. Le . . .	— 7	— 18	— 6
» » Dosde' Cen. Le . . .	— 11,5	— 14	— 11
» » Dosde' Occ. Le . . .	— 31	— 30	— 20
» » Viola Or. Le . . .	— 2	— 3	— 9
» » Viola Occ. Le. . .	— 2	— 2	— 20
» » Verva Maggiore. . .	— 5,5	— 7	—
» » Cardonne' Or. Le . . .	— 6	— 7	— 15
» » Cardonne Occ. Le . . .	— 13	— 4	— 6
» » Val Lia	— 1,5	— 2	— 4
» » delle Mine	— 6,5	— 8	— 13
» » Lago Bianco	— 7	+ 3,5	+ 3
» » Gavia	— 0,5	+ 11,5	+ 13,5
» » Sforzellina	— 14	— 11	— 4
» » Tresero' Mer. Le . . .	— 1,5	— 18	—
» » Tresero' Sett. Le . . .	— 6	— 12	— 3,5
» » Dosegu'	— 11	— 24	—
» » Vitelli	—	— 88,5 (5 ans)	— 10,5
» » Miniera	—	— 71 (5 ans)	—
» » Castelli	—	— 35,5 (3 ans)	—
» » Saent	— 3	—	—
» » Careser	— 25	—	—
» » Lamare	— 11	—	—
» » Rossa	— 7	—	—
» » Zebbru'	—	+ 6	—
» » Pisgana Occ. Le. . .	— 26,5	— 23,5	— 26
» » Pisgana Or. Le . . .	— 13	—	—
» » Adame'	+ 7	—	—
» » Salarno'	—	— 35 (2 ans)	—
» » Aviolo	— 17 (5 ans)	—	—
» » Avio	—	— 53 (4 ans)	—
» » Calotta	— 19	—	—
» » Care' Alto	— 2,5	— 25,5	— 3
» » Venerocolo	— 0,5	— 5	—
» » Niscli	— 3	— 17	— 4
» » Lares	— 0,5	— 31,5	— 9,5
» » Folgorida	—	— 65	—
» » Lobbia	— 15,5	— 7,5	—
Ghiacciaio Mandrone	— 10	— 11,5	+ 5
» » Nardis	— 8	— 10,5	— 9
» » Amola	— 6	—	— 3 (2 ans)
» » Cornisello	— 19	— 5,5	— 7
» » Presanella	— 7	— 8	— 10
<i>Alpi Orobie</i>			
Ghiacciaio Trobio	+ 8	— 11	— 12
» » Gleno	+ 8	— 7	— 6
» » Scais	+ 12	— 10	— 8
» » Parola	—	— 5 (2 ans)	—

VARIATIONS DE LONGUEUR DES GLACIERS DES ALPES ITALIENNES (Suite)

	1956/57	1957/58	1958/59
<i>Alpi Venoste Occ. li</i>			
Ghiacciaio Vallelunga . . .	— 4,5	— 9	—
» » Barbadorso di Dentro	— 23,5	— 21,5	—
» » Barbadorso di fuori	— 0,5	—	—
» » Fontana Occ. le . .	— 7,5	— 15	—
» » Saldura	—	— 8,5 (2 ans)	—
» » Ramudla	—	— 11 (2 ans)	—
» » Gogo Alto	—	— 25,5 (2 ans)	—
<i>Gruppo di Brenta</i>			
Ghiacciaio Vallesinella . . .	— 11	—	—
» » Tuckett	— 40	—	—
» » Brentei	— 13,5	+ 0,5	—
» » Sfulmeni	— 4	— 1,5	—
» » Lagol o Nardis . . .	— 4	— 27,5	—
» » Prafiori'	— 17 (2 ans)	— 2,5	—
» » XII Apostoli . . .	— 10,5 »	— 4,5	—
<i>Alpi Orientali</i>			
<i>Alpi Aurine</i>			
Ghiacciaio Quaira Bianca . .	— 3	—	—
» » Gran Pilastro . . .	— 3,5	—	—
» » Antelao Occ. le . .	— 12 (5 ans)	—	—
» » Antelao Or. le . .	— 11 » »	—	—
<i>Alpi Dolomitische</i>			
Ghiacciaio Sorapis Or. le . .	— 10,5	— 3,5	— 0,5
» » Sorapis Cen. le . .	+ 2	— 2	— 0,5
» » Sorapis Occ. le . .	— 2	— 1,5	—
Ghiacciaio Cristallo	—	— 2 (2 ans)	—
» » Popena	—	—	—
» » Cresta Bianca . . .	— 5	—	—
<i>Alpi Giulie</i>			
Ghiacciaio Canin Occ. le . .	—	—	—
» » Canin Or. le . . .	—	—	—
» » Ursic	—	— 10 (2 ans)	—
» » Montasio Occ. le . .	—	—	—
<i>Appennini</i>			
Ghiacciaio Calderone	—	—	—

Observations recueillies par M^llrs, P. Rachetto, C. Lesca, C. Socin, S. Bucchetti, A. Moretti, C. Capello, I. Cossard, C. Origlia, M. Vanni, F. de Gemini, L. Valtz, W. Monterin, S. Pignanelli, C. Saibene, I. Bellotti, A. Pollini, V. Marchetti, C. Nangeroni, A. Riccoboni, L. Ricci, L. Peretti, P. Niccoli, D. di Colbertaldo, D. Toniris.

VARIATIONS DE LONGUEUR DES GLACIERS D'EUROPE, EN MÈTRES

RÉCAPITULATION

Pays	Année	Nombre de glaciers observés	En crue	Stationnaires	En décrue
FRANCE	1956/57	7	3	—	4
	1957/58	2	—	—	2
	1958/59	3	1	—	2
SUISSE	1956/57	87	9	3	75
	1957/58	88	6	—	82
	1958/59	73	7	—	66
AUTRICHE	1956/57	41	—	3	38
	1957/58	55	1	1	53
	1958/59	43	1	1	41
ISLANDE	1956/57	29	4	—	25
	1957/58	39	3	1	35
	1958/59	31	4	1	26
SUEDE	1956/57	3	—	—	3
	1957/58	3	—	—	3
	1958/59	3	—	—	3
NORVEGE	1956/57	24	3	—	21
	1957/58	23	5	1	17
	1958/59	27	2	2	23
ITALIE	1956/57	103	9	—	94
	1957/58	94	4	—	90
	1958/59	67	4	—	63
TOTAUX					
	1956/57	294	28	6	260
	1957/58	304	19	3	282
	1958/59	247	19	4	224

CHANGES IN THE ACCUMULATION REGIME ON THE ICE CAP IN THE RUSSKAYA GAVAN AREA ON NOVAYA ZEMLYA

N.M. SVATKOV (U.S.S.R.)

Institute of Geography of the Academy of Sciences of the USSR

SUMMARY

1. The observations of the IPY II expedition established that deep glacier ice is bare on the surface of Novozemelsky Ice Cap; firn is absent. Tongues of ice cap, including Schokalsky Glacier, are existing at the expense of unregenerating ice of the central part of the ice cap. The ice cap is retreating quickly.

2. In the following 25 years Schokalsky Glacier dimensions are retreating continuously, in the first half of this period more slowly than in the second one reduced.

3. The IGY expedition stated normal firn nourishment at the altitude 570 m and more (76°N). At the iceshed between Russkaya Gavan and Blagopoluchiy bay the pit revealed the firn layer 16 m thick. The firn structure allow to consider its age as 17-20 years. The new ice is removed into the region of ablation not less than two kilometres.

4. The quantity of solid precipitation on Novaya Zemlya depends on the intensity of cyclonic activity, that had been intensified from 1930s when probably the normal glacier nourishment had been renewed. If the contemporary regime of cyclonic activity shall be maintained to 1980s, it should provide an abundant nourishment of the ice cap and lead to delay or possibly to complete cessation of the retreat of the Schokalsky Glacier.

RÉSUMÉ

1. Selon les observations de l'expédition du 2^{me} A.P.I. sous la direction de M. M. Ermolaev, l'on a établi que sur la surface du champ glaciaire de la Novaya Zemla affleure la glace appartenant à un glacier de bas-fond : le champ est dépourvu d'alimentation de névé. Les langues de décharge, y compris le glacier Shokalsky, existent au compte des glaces non-renouvelées de la partie centrale du champ. La glaciation est en état de regression rapide.

2. Au cours des 25 années ultérieures, les dimensions du glacier Shokalsky diminueront sans interruption, avec plus de lenteur au cours de la première moitié de cette période et plus vite ensuite.

3. L'expédition de l'A. G. I. a révélé une alimentation normale de névé du champ de glace de la Novaya Zemla à des altitudes excédant 570 m ($\varphi = 76^{\circ}\text{N}$). Sur le partage des eaux l'on a décapé par le travail de découverte une ascise de glace de névé et du névé de puissance atteignant 16 m. La structure de l'assise permet de croire qu'elle se forma de cours de 17 à 20 ans. La glace survenue à nouveau se déplaça à un moins de 2 km vers la région d'ablation.

4. La quantité de dépôts solides sur les glaciers de la Novaya Zemla dépend d'une activité cyclonique intense.

Cette activité se renforça dès le début des années 30, quand probablement se renouvela aussi l'alimentation normale du champ glaciaire. Le régime actuel de l'activité cyclonique se conservera, il faut le penser, jusqu'aux années 80, ce qui assurera une alimentation abondante du champ et amènera à un ralentissement, peut-être même à un achèvement complet du recul du front du glacier Shokalsky.

In the Russkaya Gavan area on the North Island of Novaya Zemlya glaciation was studied by special glaciological expeditions on two occasions—in 1932-33 and in 1957-59. During the IPY II the expedition under the leader of M. M. Yermolayev, in which K. Vvolken took part, it was established that over considerable areas of the ice cap on elevations of up to 700-800 metres there is an absence of firn nourishment and that beneath the layer of "this year's" ice on the iceshed between Russkaya Gavan and Blagopoluchiy Bay there is, in particular, blue glacial ice with a specific weight of $0.9036 \text{ gr/cm}^3(1-4)$. M. M. Yermolayev⁽³⁾ surmised that there is normal firn

nourishment only on the cupolas rising to an altitude of over 800 metres. In the lower-lying regions, the main mass of the snow is carried to the coast and out into the sea by strong winds and the remainder disappears completely during the summer. The ice sheet diminishes not only through the melting of outlet glacial tongues of the Shokalsky type in Russkaya Gavan, but also through the melting of the entire surface of the cover up to an altitude of about 800 metres. The heightened air pressure in the bubbles in the ice gave M.M. Yermolayev⁽²⁾ grounds for guessing its depth origin and its short subaerial position, and hence the extraordinary intensity of ablation. At the same time the suspended glaciers in the TSAGI and Bastion mountains indicated that the absolute altitude of the climatic snowline is 500-550 metres, while the orographic snowline comes down to sea level. Other investigators⁽⁵⁾ of North Islands of Novaya Zemlya agreed with this conclusion of M. M. Yermolayev's. The idea that glaciation on Novaya Zemlya was diminishing at a catastrophic rate became widespread⁽⁶⁾.

However, the results of investigations in other glaciation areas of the Soviet Arctic did not confirm this conclusion of the glaciological expedition of the IPYII in Russkaya Gavan and at the close of the nineteen-forties P.A. Shumsky⁽⁴⁾ found it possible, after analyzing all the factual data available at the time, to advance the supposition that the climatic snowline passes at an altitude of 400 metres.

G. A. Avsyuk⁽⁷⁾ reported that a layer of firn with numerous strata of ice totalling more than three metres in depth was discovered in the summer of 1955 in the region of the Russkaya Gavan-Blagopoluchiy Bay iceshed. An identical picture was observed in other areas of North Island of Novaya Zemlya, in spite of the fact that during his journey to Cape Zhelaniya M. M. Yermolayev^(3,4) pp. 110, 122-123) found accumulations of ice only in large depressions such as Anna Valley.

At the end of August 1957, the IGY Novaya Zemlya Glaciological Expedition of the Institute of Geography of the Academy of Science of the U.S.S.R. recorded the position of the climatic snowline on the brow of Yablonsky Barrier as 570 metres above sea level⁽⁸⁾. Its position remained unchanged in 1958 and 1959. With the aid of pits dug in the area between Yablonsky Barrier and the iceshed it was established that the thickness of the firn layer increases rapidly with a growth of altitudes. On the bed of the depression, 4.4 kilometres south-east of Yablonsky Barrier, it is approximately two metres thick, but on the iceshed it reaches a thickness of sixteen metres. V. Y. and A. B. Bazhev determined the density of the ice in a pit dug in the iceshed at the level of 770 metres (*). The results are given in Table 1.

TABLE 1

Ice density in a Pit in the Iceshed between Russkaya Gavan-Blagopoluchiy Bay

Level (m)	Specific weight (gr/cm ³)
5.10- 5.20	0.9024
9.02- 9.11	0.9019
16.65-16.75	0.9159
20.1 -20.3	0.9005
25.9 -26.3	0.9048

(*) The author is sincerely grateful to all his colleagues of the Novaya Zemlya Glaciological Expedition for their kind contributions.

The author inspected the pit and found that down to a depth of 6.5 metres large-granule firn interlies with turbid (milky) firn ice. Below that level individual strata are transparent, becoming filled with bubbles. Below 9 metres there is a predominance of stratified ice: strata of turbid ice alternate with strata of transparent ice. Nine lenslike 1-10 cm thick layers of ice were counted in the last large level of firn at a depth of 10-12 metres. Below 12 metres, ice predominates in the structure of the firn layer. The last level of large-granular, dense, compact firn of a thickness of up to 60 cm was observed at a depth of 16 metres. Beneath it is a layer of ice with clearly expressed striped turbid milky and transparent bluish bubbly ice, whose thickness steadily increases. At a depth of 20 metres the turbid milky ice forms separate thin 1-2 cm layers. Even a cursory inspection of the pit cannot fail to bring out the clearly distinguishable stable level of dense blue ice without any visible bubbles at a depth of about 17 metres. A comparison of its density with the results of the analysis of ice samples taken from beneath a layer of seasonal snow in the same area by the I. P. Y. II expedition shows that although there are differences, the ice above and below this level is not so dense. We know that as solid precipitation accumulates the density of the buried level increases under the influence of the rising pressure and cannot diminish. No other levels of ice with a density of 0.916 were found in the 26-metre pit. Without excluding the possibility that the "Yermolayev level" was in a different stratigraphic position, it evidently may be surmised that this level of congelation ice took shape on the water-tight layer of dense ice, and that this was the ice that was observed on the surface of the ice cap in 1932, while the whole layer of ice above it accumulated in the course of the next 25 years. This surmise seems to be the more correct in the light of the following facts.

While on October 23, 1932, 80 cm of snow containing 384 mm of water had accumulated on the dense blue ice of the iceshed and by the close of the winter the snow cover had grown to a height of up to 190 cm (water content—up to 770 mm), then by October 17, 1933, the height of the ice cover was already 150 cm (4). Assuming that the rate of melting in the summer and the rate of snow accumulation in the autumn of 1933 were the same as in 1932, it must be considered that at least 70 cm of snow had summered. Judging by the observations made in 1958 by E. M. Zinger, the moisture content of melting snow increases in the ablation period from 0.41 to 0.45 gr/cm³, i.e., by only 9-10 per cent. Therefore, it may be surmised that the snow that had summered in 1933 contained about 300 mm of water and that what had melted and evaporated contained approximately 470 mm of water. G. N. Yakovlev (9) has estimated that 0.1-0.6 mm of water evaporates daily in July and August in the Central Arctic. If we were even to treble this figure for the area under review and take the precipitation in the period of melting (38 mm in 1958) into consideration, then it is hardly possible that less than 440 mm of water had seeped into the glacier during the summer of 1933, because there is no run-off on the surface. The freezing of such masses of water cannot fail but to facilitate the formation of thick layers of ice and the rapid metamorphism of the firn.

Thermometric investigations conducted in 1957-1959 in the regions of nourishment and ablation at Ledorazdelnaya and Somnieniye Barrier bases showed that ice temperatures on the Novaya Zemlya ice cap are high. In the nourishment region it is several times higher than around Somnieniye Barrier, the reason for this being the great heat-insulating properties of the firn and the considerable thickness of the dense ice sheet. For an example: on November 8, 1957, a 12-ton tractor crushed a snow bridge over a crevasse in the nourishment region on an elevation of about 700 metres. The temperature of the air was close to -30°C but clouds of steam rose out of the crevasse. Wet crumbs of firn and ice rose from 10- and 15-metre levels when I. F. Khmelevskoy, O. A. Yablonsky and V. N. Genin sank a well on the iceshed in December 1957-January 1958. Subsequent temperature measurements taken by

I. F. Khmelevskoy, V. N. Genin, N. V. Davidovich, Z. M. Kanevsky, E. M. Zinger, V. V. Engelhardt and others at the levels of down to 15 metres showed that whereas in January 1958 the zero temperature was observed at a depth of 10 metres, in July it was observed below 15 metres. The minimum of -0.6°C was attained on July 29, while by August 3 it rose to zero at a depth of 15 metres. The levels lying above (from 15-10 metres) retained negative temperatures until the close of September, although early in August it rises to 1.3°C . This indicates that melt water penetrates deep along separate crevasses. The influence of melt water on the formation of ice is strikingly seen on the bed of a depression southeast of Yablonsky Barrier. Melt water running down its slopes accumulates there. The barrier prevents water from running off, and it saturates the snow cover to such an extent that it rises to the surface over an area of one and a half to two kilometres. A unique and quite impassable "swamp" is thus created. The author saw this for himself on August 13, 1957. With the appearances of cracks after melting begins, the water penetrates into the ice, the "swamp" is drained, but the water content of the remaining firn increases considerably. Firn melts much more slowly and for that reason by the end of the ablation period a smaller portion of the seasonal cover melts than on the slopes or the watershed. The formation of whitish bubble ice thus increases "artificially" at the bed of the depression and that is where it moves farther in the area of ablation than on the slopes of the depression.

The region of the iced bed where the pit was dug is situated at the foot on the gently rising slopes of two neighbouring caps and it is quite possible that an inner flow of melt water comes from there. This is indicated both by the zero temperature at a depth below 10 metres in January and the wet crumbs of firn and ice at the 10- and 15-metre levels. In other words, melt water can stimulate the accumulation of ice on the iced bed as well.

Table 1 shows that at a depth of 20 metres the density of the ice is smaller than at the depth of 5-9 metres. From this we may draw the conclusion that the formation of this level proceeded with a somewhat lesser participation of melt water, whose quantity diminished evidently through the influence of lower summer temperatures. Conversely, the formation of the level of dense blue ice most likely proceeded under the influence of abundant melt water, in other words, during the period of warm weather.

The features of the accumulation of solid precipitation on Novaya Zemlya and indeed, in other regions of present-day glaciation, depends not only on the latitude of the region but also on the peculiarities of the atmospheric circulation and fluctuations over a long period due to changes in solar activity. N. I. Tyabin⁽¹⁰⁾ proved the very close dependence of the different types of atmospheric circulation, and through this showed that the hydrological situation in the Atlantic regions of the Arctic depends on solar activity as expressed by the maximum area of each group of spots. The first half of the present century is characterized by a diminution of the latitudinal and an increase of the meridional transportation of air against a background of rising solar activity. In this connection, the quantity of water entering the Gulfstream is increasing, the temperature of the water on the Kola meridian is rising and a "chain reaction" of the whole hydrological situation in the Barents Sea is taking shape: the ice is decreasing, water and local masses of air are growing warmer and the water content of the air is increasing. These factors tended to increase facilitate the settlement of moisture on the glaciers. Nevertheless, hardly any importance can be attached to the participation of moisture of local Arctic region in the nourishment of the glaciers of Novaya Zemlya. Most of the solid precipitation is brought to these glaciers by air masses of southern origin, primarily from the Atlantic, and, consequently, depends on the type of cyclonic activity predominating in autumn, winter and spring, because during the ablation period precipitation at elevations of up to 800 metres is almost exclusively liquid.

The synoptic charts of the northern hemisphere for 1899-1954 have been analyzed by B.L. Dzerdzeyevsky (¹¹). He has singled out three types of circulation: latitudinal, meridional and irregular latitudinal, and determined the frequency of each type for every point of the hemisphere in January, June and September in the shape of averages or ten-year running averages. The results of this analysis are thus quite acceptable for an interpretation of the phenomena taking place on the ice masses of Novaya Zemlya as well. B.L. Dzerdzeyevsky drew the conclusion that in January during 1931-1940 the frequency of the air transportation along parallels dropped by approximately 12 per cent and that the inter-latitudinal exchange increased correspondingly. This feature of the atmospheric circulation in January was expressed with particular force in the second half of the decade, when precipitation must have been intensive. On the other hand, a weakening of the latitudinal movement of air masses in the summer period began ten years earlier and achieved its development (20 per cent) by the mid-thirties (the ten year period covering 1929-1938). In the autumn months the latitudinal movement weakened substantially in 1926-1940, while the meridional movement reached its maximum. The annual mean for the period of 1932-1941 is distinguished by a minimum frequency of latitudinal movement of air, a minimum frequency of latitudinal movement with separate invasions of southern cyclones, and a maximum interlatitudinal exchange. In the preceding decades there was a high frequency of invasions of warm masses of air in the summer and an insufficiency of such masses of air during the cold period of the year. In the course of the 1920s there was thus less-than-normal precipitation and high temperatures during melting in the region of nourishment of the Novaya Zemlya glacier, i.e., the conditions were created for the melting not only of winter precipitation but also of the surface levels of the ice sheet. In those years the situation was possibly the same as was observed in 1958-59, when 152 cm of snow containing $\cong 600$ mm of water accumulated on the watershed in winter, and not only all this snow but also a 20 cm residue of the previous year's snow with 90.3 mm of water melted in the summer.

As the winter latitudinal movement dropped and the meridional movement increased (by the nineteen-thirties), the annual total precipitation rose through the fall of solid precipitation. An equilibrium between accumulation and ablation rose on the Russkaya Gavan meridian and on elevations of about 700-800 metres evidently in the early thirties and subsequently rapidly fell to approximately 130-200 metres. As observations during the IPYII showed, the accumulation of solid material resumed already in 1933. At the same time the drop in the accumulation of snow in 1957-59 compared with 1932-33 and the intensity of melting seems to indicate a change in the course of the circulation (as B.L. Dzerdzeyevsky surmises), i.e., we may in the immediate future expect an intensification of the metamorphism of solid precipitation and even changes in the accumulation regime—a transition from snow nourishment with the participation of melt water to ice nourishment with a drop in the quantity of precipitation.

Thus, with the regard to the nourishment of the ice cap of Novaya Zemlya it seems possible to speak of two changes in the intensity and features of the onset and transformation of precipitation. In the late nineteen-twenties there was an ice type of nourishment, which in the early nineteen-thirties gave way at the predominance of ablation over accumulation at the elevations of up to 800 metres. Nourishment of the ice then resumed in the shape of accumulations of firn in the mid-thirties with the participation of melt water, and the intensity of nourishment remained high right up to the mid-fifties. Then the quantity of precipitation dropped and now there again are indications of a transition to the ice type of nourishment with short-lived equilibria between the income and expenditure of matter on these elevations as a consequence. It is possible that, as was the case in the early nineteen-thirties, ablation will

become predominant for a short period. It is important to state that contrary to M. M. Yermolayev's surmises no catastrophic reduction of the ice sheet of Novaya Zemlya is taking place. There is a fluctuation in the intensity and type of nourishment and this is closely dependent on the fluctuation of atmospheric circulation throughout the world.

The above named features of the nourishment in the Novaya Zemlya ice cap are reflected in the dynamics of the Shokalsky Glacier. Although between IPY II and the IGY no special observations were made of the position of its front, we have some data about its dynamics. By comparing these data we can see that the front did not retreat continuously. In the period between 1933 and 1952 it was either stable or even tended to move somewhat forward, because in 1952 the front was somewhat closer to the sea level than in 1933. Judging by the extension of the right border valley, the rapid retreat commenced at some period between 1944 and 1952 (¹²), while in the years between 1952 and 1959 (*) its retreat reached 450-460 metres. Shokalsky Glacier

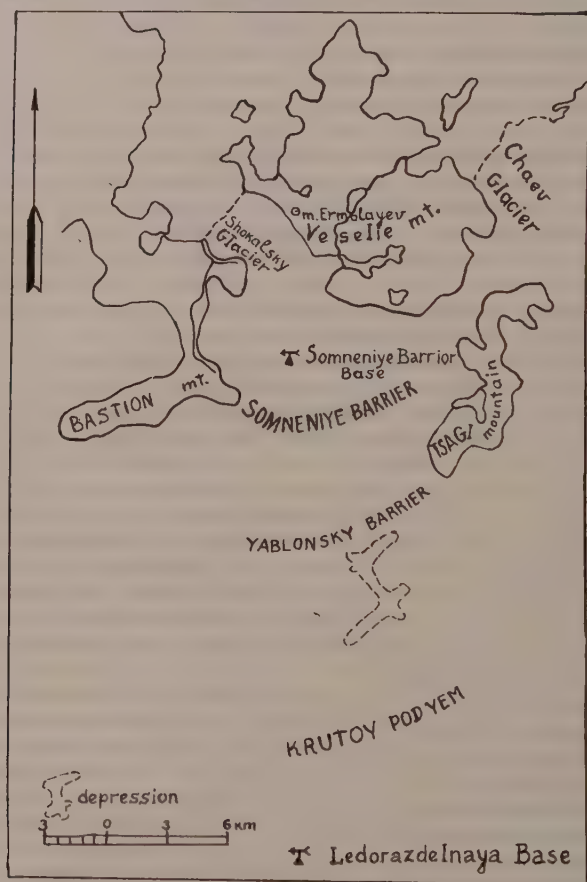


Fig. 1 — Region of observations made by the Novaya Zemlya Glaciological Expedition.

(*) In 1959 the position of the front was recorded by V. S. Koryakin.



Fig. 2 — Size of the Shokalsky Glacier in 1933, 1952, 1959.

1. Coast line.
2. Front in 1933.
3. Front in 1952.
4. Front in 1959.

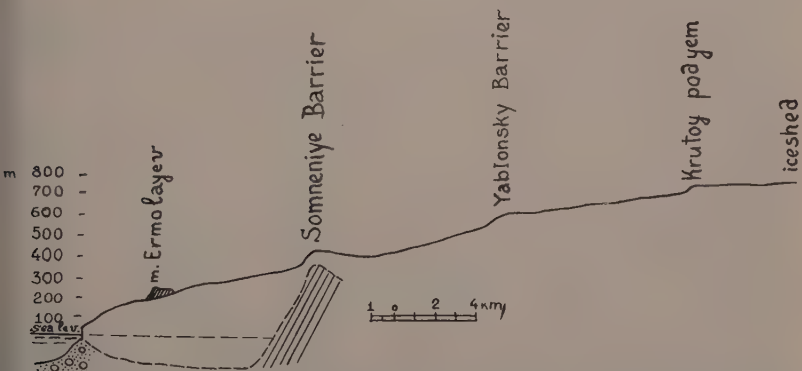


Fig. 3 — Cross-section of the North Island ice cap of Novaya Zemlya in the area of Russkaya Gavan.

became the only outlet stream of the Novaya Zemlya ice cap on the northern coast of the island that did not retreat in 1933-1952. The depth measurements taken by V.S. Koryakin along the front of the glacier and the seismic soundings made by K. Vyolken in combination with the above-mentioned features in the position of the front give us the grounds for surmising that this glacier is not floating and for that reason is less sensitive to changes in climatic conditions. The rapid retreat of the front, after 1952 in any case, possibly reflects the fall in intensity of nourishment in the period from the close of the twenties to the beginning of the thirties in combination with the general warmer climate in the second quarter of the present century. In view of the more intensive nourishment that began in the thirties and the colder climate that began in the mid-fifties, it is quite probable that we may expect the retreat of the front of the Shokalsky Glacier to slow down in the future.

In spite of the tendency of the glaciation of the globe to drop, nourishment of glaciers has improved in recent years in many mountainous countries. Not only has it been noted that the mass of some glaciers has increased but that they are expanding in the Caucasus ⁽¹³⁾, in Montana ⁽¹⁴⁾, in the Alps ⁽¹⁵⁾ and in some other areas. The opinion is advanced that the recent retreat of the glaciation in Antarctica has ceased ⁽¹⁶⁾. This points to the world character of the phenomenon we are speaking of and to the great sensitivity of glaciers to climatic fluctuations, and enables us once again to convince ourselves of the high degree of equilibrium between natural processes and the immense mobility of this equilibrium.

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METEOROLOGICAL FACTORS IN THE GREAT GLACIER ADVANCE (1690-1720)

G. MANLEY (Great Britain)

SUMMARY

Evidence is provided for a well-defined short-term climatic deterioration in the last decade of the 17th century, based on some recently-discovered primitive instrumental-meteorological observations kept at London. Notably from 1692 to 1701 the predominance of cold springs and cool unsettled summers can be related to the onset of the very marked Alpine glacier advance which ended about 1715. These seasons were associated with bad ice years off the coast of Iceland. The circumstances giving rise to such a notable climatic episode require further investigation.

RÉSUMÉ

La récente découverte des résultats d'observations météorologiques à l'aide d'instruments primitifs à Londres à la fin du 17^e siècle a fait constater qu'il s'y produisit une détérioration du climat, notable entre 1692-1701, caractérisée par une prédominance de printemps froids et d'étés frais et par conséquent une grande extension des glaciers alpins jusque 1715. Ces périodes sont aussi associées avec l'apparition de glaces flottantes sur les côtes d'Islande. Les circonstances qui accompagnèrent une telle courte récession climatique demandent encore des investigations.

It appears worth while to try to establish the reasons for the behaviour of European glaciers in the past. The object of this paper is to summarise results derived from the reduction of some primitive English meteorological observations, taken in and near London, before 1706; and to relate them to the behaviour of the European mountain glaciers. These early English meteorological journals are, in part, a very recent discovery among the MSS in the Bodleian Library at Oxford. They cover a critical period: 1680-1700, which was followed by a very decided advance of the glaciers in the northern and western Alps, and also by decided advance in Iceland; culminating, in both regions, about 1715-1720.

This short but distinct climatic recession towards the end of the 17th century was marked not only by glacier advance, but also by considerable distress and economic disturbance in northern lands. Hence it has significance for a great part of Europe. Contemporary meteorological journals are very few; hence it seemed that these preliminary findings should be described here, as a possible stimulus to discussion.

We have, in the London area, daily observations of temperature for 1680-1699 and 1699-1706, and also some scattered short records before that. There is also an interrupted series of readings covering about two-thirds of the period 1692-1703. Together with these are daily observations of the weather, in some detail, testifying to a keen observer, and apparently homogeneous throughout, from late in 1668 to the end of 1700. Two other daily records cover 1697-1706 and 1699-1717. There are also several shorter records. Wind directions are recorded daily from May 1670 onward.

The journals so far investigated are sufficiently detailed to enable us to count up the number of days with precipitation observed to fall, and also the number of days with snow or sleet falling. The standards of observations approximate to those of a well-maintained climatological observing station to-day. It is probable that «the number of days with precipitation observed to fall» will be very slightly lower than «the number of days with measurable rain» (i.e. 0.2 mm or more) recorded at present-day official stations, because the seventeenth-century observer might now and then fail to record a «day of rain» resulting from a short shower falling during the night hours. But the difference is not likely to be large.

Accordingly, the fact that the average annual number of days with «precipitation observed to fall» was 161 over the twenty years 1671-1690, testifies to a regime differing little from that of to-day. Assembling the data from three London stations (Kew, Greenwich, St. James' Park) for 1881-1915, the average is 164. The distribution by months is as follows the figures being smoothed to 0.5 :

	J	F	M	A	M	J	J	A	S	O	N	D	
1671-1690	13	11	15	15	12.5	14	14.5	14.5	12	14.5	13	11.5	161
1881-1915	15	13	14	12	13	12	13	13	11	16	15	16	164
1901-1930	17	14	15	13	12	11	13	14	11	16	16	17	170 (approx., based on maps)

From this it appears that there is no serious difference, although winters appear to have been relatively less rainy in the earlier period.

What becomes noticeable at once, however, is the difference in the figures for the decade 1691-1700 compared with the preceding period:—

	J	F	M	A	M	J	J	A	S	O	N	D	
1691-1700	13	15	15	17	17	15	14	16.5	14	13	14	16	179

In this decade the average number of days with precipitation observed to fall increased by 11% compared with the earlier period 1671-1690, and the increase was most notable in the spring and summer months. The conclusion is inescapable that the majority of the summers were appreciably cooler and more unsettled. The effects in the spring months will be discussed later. For 1701-10 the approximate average is 168.

The second point to notice with regard to this decade is the very marked increase in the average annual number of days with snow observed to fall. Over the decade 1671-1680 the average was 17.1; in the decade 1681-1690 the average was 18.3; but during 1691-1700 the average rose to 25.5. For 1701-1710 the average is 16.0.

It is pertinent to compare these figures with the average characteristic of good climatological stations to-day, which for 1921-1950 is approximately 13 in the London area. In the later decades of the nineteenth century averages of 16-17 are found, and therefore it does not appear that the period 1671-1690 differed seriously in character from say, 1871-1890. Within the period 1671-1690 there was one excessively severe and prolonged winter (1684) while 1679, 1681, 1685 and 1689 were all distinctly cold; on the other hand, 1671, 1676, 1680, 1682 and especially 1686 were mild. The March of 1674 was outstandingly cold and snowy; that of 1688 was also cold.

We may now turn to the results derived from the very primitive thermometers. These are extremely difficult to interpret on account of their many imperfections, but the best reconstruction I can put on them, largely based on the overall control provided by the snow statistics, provides a series of monthly mean values from 1680 onward. These point to a fall in the mean annual temperature for the decade 1691-1700 of about 0.6°C below that of 1681-1690. In the individual seasons the decline ranges from 0.8°C in winter, 1.0° in spring, 0.2 summer, 0.4 autumn. The marked decline in the spring months (March to May) deserves our attention.

1. GOLD SPRINGS AND THE ADVANCE OF GLACIERS

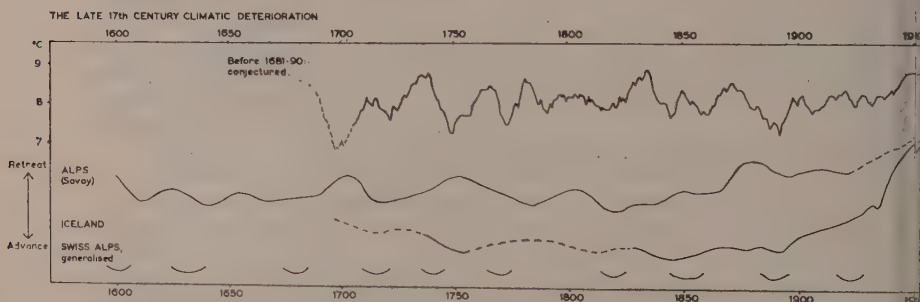
In temperate latitudes glacier behaviour has been shown to be related primarily to variations in the mean temperature of the ablation season and secondarily to variations in the precipitation, which in turn affect the accumulation. At the critical

altitudes to which the smaller Alpine and Scandinavian glaciers descend a quite close relationship exists between the accumulated temperature above 0°C during the warmer months, and the advance or retreat of the snout. This has been shown by Th. Zingg for the Turtmann glacier, whose behaviour lags by about nine years on the variations in the five-year total of accumulated warmth (U.G.G.I. Rapports, Commission des Neiges et Glaces, Brussels meeting, 1951, pp. 266-269).

At these altitudes where we may expect that the mean annual temperature will be slightly below 0°C , it will evidently be in the months of April and May, and again in October, that the average temperature will vary on either side of 0° . In a warm spring much of the precipitation will fall as rain rather than snow; but throughout a cold May, at 2500 m in the Alps or 1500 m in Norway, practically all that falls will still be snow. If April and May are not only cold, but also cloudy and unsettled, the total accumulation may thus easily attain 50% above that of a normal winter season, and in a cloudy cold May the beginning of ablation will be the more delayed. Moreover, in northern and western Switzerland and Savoy, the snowfall in March makes in a normal winter an important contribution to the accumulation. In eastern England a decidedly cold March is commonly accompanied by considerable snowfall, largely on account of the instability generated in Arctic air approaching across the sea; and on meteorological grounds it would appear that such Marches would likewise give rise to considerable snowfall further south.

Hence the predominance of temperatures below the average during the three months March-May, in England, would seem likely to be associated with a decided increase in Alpine accumulation in one or more of these months at these critical altitudes; and with little or no ablation. We might therefore expect that the fluctuations of the mean temperature of March-May in England, when smoothed into five- or ten-year running means, would be closely related to the behaviour of the outer Alpine glaciers. A reasonable association with the fluctuations of Icelandic glaciers might also be expected. With Scandinavia the association might be less apparent, on account of the fact that east and north-east winds in Central Norway are relatively dry.

These facts are assembled in the diagram below. The mean temperatures for the London area, deduced from the early observations, have been adjusted to form an approximate continuation of the «Central England» series of monthly means (Manley, *Arkiv für Met. u. Bioklim.*, 1959, pp. 413-433). The curves have been redrawn (from Ahlmann, «Glacier Variations and Climatic Fluctuations», *Amer. Geogr. Soc.*, N.Y., 1953). They derive from Mougín's discussion of the glaciers of Savoy and Thorarinsson's discussion for the Vatnajökull glaciers in Iceland. Generalised dates of the forward positions of the Swiss glaciers are also shown; these are based on Charlesworth (The Quaternary Era, London, 1957).



OSCILLATIONS of the MEAN TEMPERATURE of the SPRING MONTHS (March-May) in ENGLAND and the behaviour of GLACIERS (Alps, Iceland)

TEMPERATURE 10-year moving averages GLACIER behaviour based on AHLMANN (1953) and CHARLESWORTH (1957), summarising earlier authorities.

It will be observed at once that the decade 1691-1700 represents a short but very notable climatic recession. In particular, the prolonged cold and snowy winter of 1695, followed in England by a decidedly cool unsettled summer, should fairly be linked with the worst ice year on the coasts of Iceland of which we have knowledge (cf. Koch, *Medd. om Grönland*, 130, 1945). The marked increase about this time in the amount of ice in the East Greenland Seas is noteworthy; and in 1695 ice was observed all round the coasts of Iceland. The decidedly sharp advance of the Alpine glaciers is evidently in large measure a consequence of this marked fall of the spring temperatures, associated with cool and unsettled weather both in spring and summer, the «lag» being of the order of ten or twelve years. It may be noted that the autumn temperatures also fell; but the increased frequency of unsettled cold weather in the spring was probably more significant.

The reasons for this sharp fall in the overall averages of temperature, beginning about 1688-1690, remain to be discussed. It was characterised by all the factors making for a decided increase in accumulation and decreased ablation in the majority of years, until about 1702-1703. The course of the curve suggests that no conspicuous glacier advance appears to have developed unless the mean spring temperature in England over a decade has fallen to levels below 8.3°C. Although a slight fall can be seen in recent years (since 1950) this value has not yet been passed. It remains to be seen whether this criterion will hold in future.

The fact that the spring temperatures, which are primarily affected by the temperature of the Atlantic ocean off W. Europe, have not fallen to the levels formerly prevailing appears worthy of further examination in relation to the known recession of the Arctic sea-ice. It may be noted that the fluctuations in the mean temperature for the whole year, as shown by decadal running means, also occur synchronously with those of the spring months. This in turn suggests that the water of the N.E. Atlantic as a whole has become warmer, and that this is a principal reason for the maintenance of the present amelioration. This goes far to support the view that a principal factor governing the behaviour of the glacier of W. Europe lies in the temperature of the Atlantic waters and the relative breadth of the colder surface waters lying to the north-west; again, the ultimate cause of these changes for discussion. Such deductions as we can make regarding the behaviour of the glaciers in the Late-Glacial in N. W. Europe (the Post-Alleröd, or «Younger Dryas» period) lead to similar conclusions (Manley, *Geogr. J.*, 1951). But changes in the surface temperature of the N. Atlantic cannot be the only factor, as Heusser (*Ecological Monographs*, 26, pp. 263-302, October 1956) and others have shown that glacier variations in the Canadian Rockies show a broad agreement with those of Europe. We must await with interest any findings from the Altai glaciers in south-central Siberia, before we can decide what additional part is played by the waters of the North Atlantic, beyond that which can be ascribed to the vicissitudes of the upper-westerly flow and the temporary displacements of the relative position of «ridges» and «troughs» in the wave pattern that the upper westerlies display. The relatively rapid European development of the «climatic recession» of the late 17th century is a matter of interest. Whether its effects were largely peripheral to the North Atlantic, remains for glaciologists in other parts of the world to elucidate.

A more extended discussion of these 17th century records is in preparation.

OBSERVATIONS SUR LE RECENT RETRAIT EXCEPTIONNEL DE CERTAINS GLACIERS DANS LES ALPES OCCIDENTALES PIEMONTAISES

Luigi JERETTI (Italie)

RÉSUMÉ

L'Auteur signale le retrait exceptionnel, mesuré dans ces dernières années — pendant la phase actuelle de recul général — aux fronts de quelques glaciers, grands ou petits, des Alpes Occidentales piémontaises.

En comparaison aux oscillations négatives moyennes annuelles d'une dizaine de mètres, les retraits dont il est question ont été, pour une seule fois, de l'ordre d'une centaine de mètres ou encore plus; après quoi le recul a continué de nouveau avec le rythme normal.

Le phénomène, en général, s'accomplit avec l'ablation totale d'un trait intermédiaire de la langue, en amont du front; il en résulte, quelquefois, une plaque isolée de glace « fossile ».

Le raccourcissement de la langue n'est pas une simple coupure : on doit l'interpréter comme étant la conséquence du rétablissement de l'équilibre altimétrique du glacier en relation avec de nouvelles conditions d'alimentation du bassin dissipateur, après une période prolongée de ralentissement des oscillations frontales, provoquées par des contingences locales.

SUMMARY

The author indicates cases of exceptional regression of the front part of some big or small glaciers of the Western Piedmontese Alps, recorded in these last years during the present phase of general regression.

In comparison with annual average negative oscillations in these same glaciers of a few metres, said withdraws have been, now and then, of some hundred metres, or even more, and, afterwards, the regression has gone on with normal rhythm.

Generally the phenomenon has occurred with a total suppression of an intermediary part of the strip upstream of the front part, and sometimes there remained a large plate isolated of ice.

The said glaciers retrenchement is not due to a simple mechanical cropping but must be interpreted as the effect of altimetric equilibrium reestablished on the glacial mass in reference to modified feeding conditions of the dissipating basin after a long period of relative inertia of frontal oscillations due to local contingencies.

Dans ces dernières années on a observé avec une remarquable fréquence — sans qu'ils aient été jusqu'aujourd'hui sujet d'étude spéciale — des retraits annuels tout à fait anormaux aux extrémités de quelques glaciers, dont les langues s'étendent sur une pente rocheuse ou détritique, plus ou moins régulièrement inclinée, ce qui va exclure les cas de tronquement par glissement, éboulement, etc.

Sans aucune variations préalable du rythme du retrait, l'extrémité du glacier après un an, apparaît reculée d'une longueur — en projection horizontale — plusieurs fois multiple du recul mesuré dans chacun des ans précédents et suivants. La réduction de longueur est naturellement liée à une correspondante élévation du niveau frontal du glacier.

Le phénomène doit être envisagé dans le cadre du retrait général des glaciers des Alpes italiennes, qui se prolonge, presque sans interruption, depuis 1890. Dès lors les principaux glaciers du versant piémontais ont reculé annuellement d'une longueur, en moyenne, d'une dizaine de mètres. Les retraits annuels extraordinaires dont fait mention cette note, sont de l'ordre d'à peu près une centaine jusqu'à quelques centaines de mètres.

Les exemples les plus significatifs — entre ceux qui ont été contrôlés directement par l'Auteur avec des mesures périodiques et des levés topographiques — concernent deux parmi les plus grands glaciers de la Vallée d'Aoste.

Le *Glacier de Grand Croux* (2 km²), dans le Massif du Grand Paradis, occupe avec son bassin d'alimentation l'amphithéâtre terminal de la Vallée de Nontey, repartie en trois coulées, qui confluaient dans le bassin dissipateur et se réunissaient dans une langue étroite, descendant en direction du Nord dans l'auge de la vallée axiale : son extrémité en 1931 touchait la cote de 2244 mètres s.l.m.

Depuis peu d'années les coulées orientale et centrale se sont séparées dans leur partie inférieure, en abondissant sur deux replats suspendus comme des balcons au dessus de la coulée occidentale.

Celle-ci, qui n'était plus alimentée du côté droit, recula régulièrement jusqu'à l'été 1953, de 6-10 m par an, tandis que sa cote frontale remontait de 1-2 m par an. En 1953 la langue, recouverte entièrement d'une couche de moraine à gros blocs et terminée par un raide talus à contour triangulaire, se rattachait encore au corps du glacier, mais se présentait déjà coupée par un système de larges crevasses transversales avec des effondrements superficiels dans une zone en amont du front.

Dans l'été 1954 on remarqua la langue interrompue par un large affleurement du lit rocheux dans la zone des crevasses. Le nouveau front du glacier s'était déplacé d'environ 250 m en amont et de 110 m plus en haut (à cote 2400). Le bord rectiligne, abouissant en angle subtile, a poursuivi son retrait moyen, mesuré dans les étés suivants.

L'ancienne langue, entamée du côté en amont par un grand creux d'effondrement et de fusion locale, est maintenant réduite à une plaque étendue et épaisse de glace «fossile».

A peu près les mêmes conditions se repètent pour le large cirque au bout de la Valpelline dans le groupe orographique Tête Blanche — Dent d'Hérens, d'où le *Glacier de Tza de Tzan* (4 km²) descend en direction du Sud. A sa langue confluaient du côté gauche la langue du Glacier des Grandes Murailles, étalé sur un haut replat par presque 7 km² de surface. La coulée de glace avançait dans la vallée, en 1934, jusqu'à la cote 2245.

Encore en 1952 les deux langues semblaient se souder latéralement au dessous d'une bande de moraine moyenne. En 1952 la langue du Glacier axial de Tza de Tzan s'était complètement individualisée, très mince et subtile, et aboutissant à la cote 2500 (à peu près), presque 600 m en amont du bord de la langue plus avancée du Glacier des Grandes Murailles, presque 1,5 km en amont de la position frontale en 1934. Même dans la zone entre les deux langues on observait de grands blocs de glace isolés avec des creux de dissolution.

Il est à prévoir un rapide retrait à venir de la langue aussi du Glacier des Grandes Murailles.

Parmi les petits glaciers l'Auteur a signalé déjà le cas du *Glacier de l'Agnello* (Vallée de la Doire Ripaire, Massif d'Ambin) : la nappe glaciaire (1 km²) qui en 1929 descendait continue au NO, en 1933 devient tout à fait séparée par l'émergence graduelle d'un plat saillant rocheux transversal en deux larges bandes superposées, dont l'inférieure (Glacier septentrional de l'Agnello), blottie dans un sillon étroit, se divisa à son tour dans le long en deux tronçons, l'un étendu complètement au dessous de la limite des neiges et tous les deux maintenant disparus. De semblables modifications de moindres glaciers aplatis, qui dépassent de peu, d'en haut et d'en bas la limite locale des neiges, ont été remarquées assez fréquemment dans ces dernières vingt ans, trouvant leur origine, chaque fois, des brusques et exceptionnels retraits aux fronts des tronçons supérieurs qui ont évolué ensuite de manière normale.

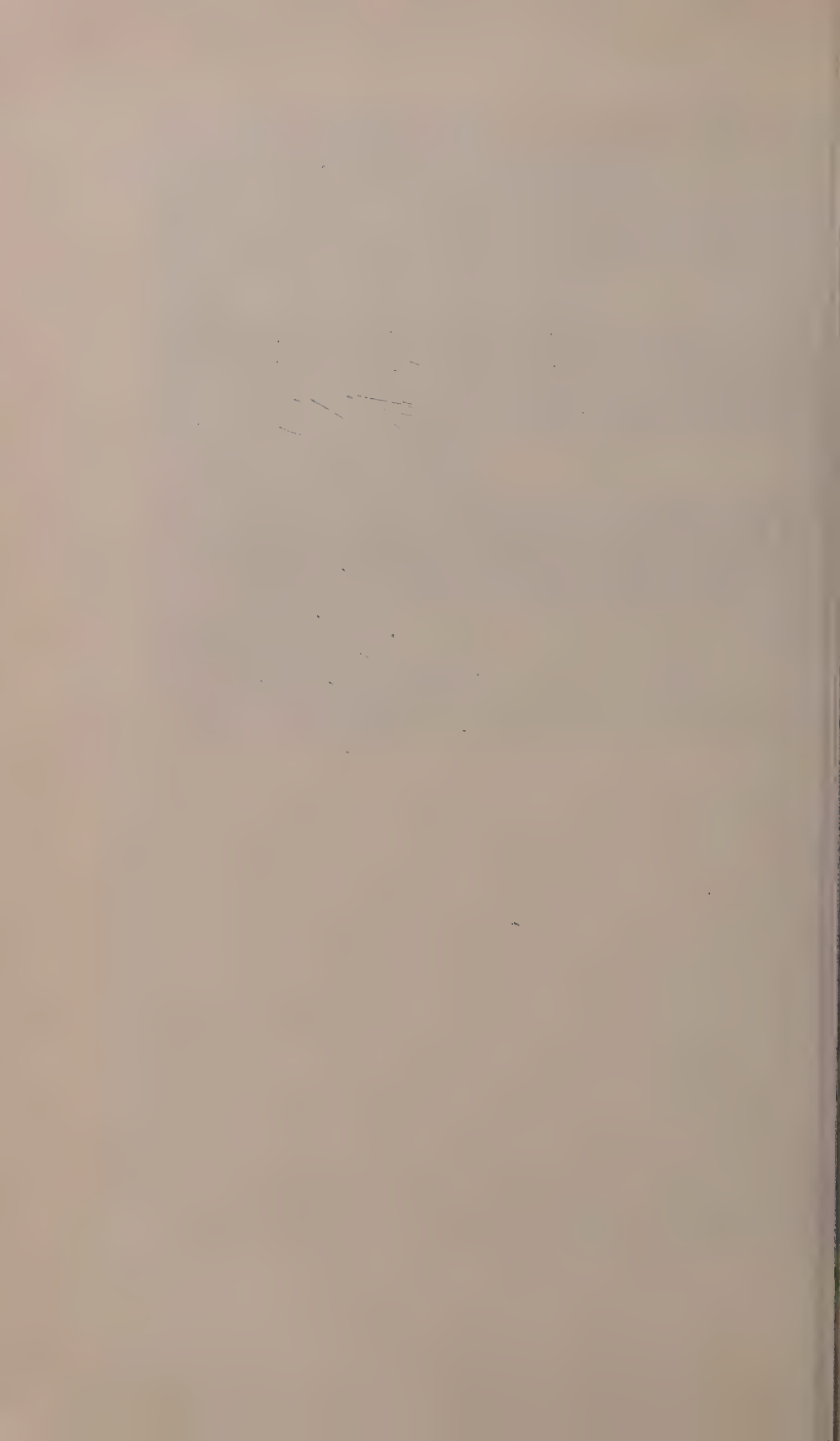
L'interprétation, en général, de pareilles phénomènes paraît assez simple, quoique elle soit susceptible de précisions ultérieures pour chaque cas considéré.

A la variation altimétrique de la limite locale des neiges, conséquente à une période prolongée — telle que l'actuelle — de conditions climatiques favorables à l'ablation glaciaire, correspond le retrait frontal des langues, même avec un certain hystérésis. Dans des conditions morphologiques particulières (langue abritée de l'insolation par son exposition ou par une couverture détritique, etc.), peut se réaliser — ainsi que pour les glaciers mentionnés — une presque inertie des oscillations du front glaciaire : alors il ne représente plus la situation variable d'équilibre entre l'alimentation par la glace qui s'écoule d'en amont et l'ablation qui agit plus intensément en aval.

Toutefois, l'équilibre physique et mécanique du glacier va se rétablir, et d'autant plus rigoureusement si la décrue générale du glacier a déterminé une forte réduction dans l'alimentation du dissipateur, comme précisément lorsque il lui vient à manquer l'apport d'une coulée confluyente. Là où l'épaisseur de la langue est réduite et son lit bien peu incliné, la vitesse d'écoulement de la glace peut presque s'anéantir. Alors l'ablation locale y procède uniformément de la surface au fond, plutôt que d'en aval vers en amont, jusqu'à provoquer la fusion totale de la glace dans les endroits de la langue où son épaisseur est moindre.

Il s'en suit l'interruption transversale de la langue, ou sa division avec des résidus «fossiles» : le vrai front du glacier, de nouveau en condition d'équilibre, va brusquement se déplacer beaucoup en amont et en haut, avec un soudain retrait de grandeur métrique anormale auquel ne correspond absolument aucune variation brusque dans le régime des facteurs climatiques. Autrefois, on observe pareillement quelques uns des caractères morphologiques et des conditions mécaniques cités auparavant : ils déterminent encore des reculs anormaux des glaciers, en conséquence d'un rétablissement accéléré de leur équilibre; mais le phénomène se poursuit et se répartit en plusieurs années : c'est le cas de deux autres grands glaciers de la Vallée d'Aoste; le *Glacier du Ruitor*, dans le Groupe des Miravidi, et le *Glacier de la Brenva*, dans le Massif du Mont Blanc, pour lequel la crue, précédant la décrue actuelle, avait elle aussi un caractère d'événement extraordinaire et une allure anormalement accélérée.

INFLUENCE DU CLIMAT SUR LES GLACIERS
RESPONSE OF GLACIERS TO CLIMATE



THE INFLUENCE OF CLIMATIC VARIATIONS ON GLACIERS

J.F. NYE

H. H. Wills Physics Laboratory, University of Bristol

SUMMARY

The paper describes, in non-mathematical terms, a recent theory on the response of glaciers to changes in rate of accumulation and ablation. The changes considered may either be the ordinary winter-summer changes, or they may be of longer period and so represent variations in climate. Temperature changes of the ice are not considered. There are only two essential assumptions: that the ice is incompressible, and that the discharge at any given cross-section of a glacier is, to a sufficient approximation, decided by the level and the slope of the upper surface. It then follows: (1) that the effects of climatic change are propagated down a glacier by kinematic waves; (2) that, in a certain sense, the lower regions of a glacier are inherently unstable; and (3), following from this, that the snout of a glacier is extremely sensitive to climatic change. Graphical examples are shown in which wave propagation and instability are clearly seen. Diffusion of kinematic waves is mentioned briefly, and the paper goes on to describe the response of a simple glacier to a cyclic change in rate of accumulation (one Fourier component of the complicated change that actually occurs). The lower part of the glacier shows not only a direct response to the cyclic change, but also a delayed response, due to the arrival of kinematic waves from upstream.

The theory enables one to calculate, in principle, the response of any glacier to any change in rate of accumulation. The necessary measurements which should be made for testing the theory are outlined.

1. INTRODUCTION

The question to be discussed is: how, in detail, does a glacier or an ice-sheet respond to a change in the rate of accumulation and ablation? We have in mind (1) the periodic changes caused by winter and summer, (2) the fluctuations of net accumulation and ablation from year to year, and (3) the slower variations due to gradual changes in the prevailing climate. Any associated changes in the temperature of the ice will be ignored; we shall be concerned only with changes in accumulation and ablation. The basic mathematical theory applicable to this situation has recently been given (Nye 1960); by its use one can calculate, in principle, the response of any glacier to any change of accumulation and ablation. The object of the present paper is to outline the salient features of the theory in non-mathematical terms, to emphasise the assumptions and the conclusions, and to suggest what measurements should be undertaken to test it. It is hoped in this way to help the discussions that are planned to take place on glaciers and climate during the conference.

2. MAIN FEATURES OF THE THEORY

The theory is formulated for the general case of a glacier in a valley of varying width, but it is easier to begin with the simpler case of flow down a valley of uniform rectangular cross-section of unit width. Let us consider a particular cross-section at a particular time. We define two basic quantities: the discharge q (sometimes called the «flow»), which is the volume of ice passing through the section in unit time; and h , the thickness of ice at the section. We then assume that at each cross-section there is, to a sufficient approximation, a functional relationship between q and h . This

may be something like a fourth-power relationship, $q \propto h^4$; but it is not necessary to make any definite assumption about the form of the relation; it is enough that, to a sufficient approximation, some relationship exists. It is assumed that the ice is incompressible, which is a good approximation except in the upper layers of the firn region, and, to begin with, we ignore accumulation and ablation. Then, without any further assumptions, it follows that there will be kinematic waves which move down the glacier with velocity c given by

$$c = \frac{dq}{dh}.$$

If $q \propto h^4$, the wave velocity is 4 times the forward velocity of the ice itself, averaged over the section under consideration.

We use the phrase «kinematic wave» here to mean, not a moving wave-form, but simply a moving point for which q is constant. The point moves down the glacier at a speed c different from that of the ice, and we speak of waves of constant q . Such waves are not as easy to visualise as waves of constant thickness. One must imagine an observer who is situated on the glacier at a particular point and a particular time. He makes a note of the instantaneous value of the discharge through the section where he happens to be. He then moves down the glacier at the speed c , which will, of course, vary down the length of the glacier. As he moves he notes the discharge at the successive sections on which he stands. He will then find, according to our result, that all the discharges he has noted down are the same. We say that he has remained on a wave of constant q . These kinematic waves are the agencies by which the effects of climatic change are transmitted down the glacier; they play the key role in the mechanism by which a glacier responds to climatic change.

The above discussion ignores the effect of accumulation and ablation. When this is taken into account we find that it modifies, but does not destroy, the wave property. As one moves at velocity c , q is no longer constant, but changes at the rate a per unit distance, where a is the rate of accumulation (negative for ablation).

Next we introduce the idea of a steady state. If the rate of accumulation and ablation, which of course varies continuously down the glacier, were constant for a long time, the glacier would eventually reach a steady state. Of course, such a steady state never occurs in practice because of winter-summer and longer-term changes, but nevertheless it is instructive to consider it as a hypothetical possibility. In such a steady state each point (cross-section) of the glacier will have a certain thickness, and there will be a certain discharge through it. Each point will also have associated with it a certain wave velocity c_0 (for most points this roughly 4 times the ice velocity, but at the snout it is equal to the ice velocity). The physical significance of c_0 is that it represents the velocity with which small disturbances from the steady state are propagated down the glacier.

We now enquire what will happen if a thin uniform layer of the glacier is suddenly melted away, so that the glacier is now no longer in equilibrium with the rate of accumulation and ablation. The question can be answered without making any further assumptions. At places where the wave velocity c_0 increases as one goes down glacier (dc_0/dx positive, where x is the distance down the glacier), it can be shown that the glacier surface begins to return exponentially to its original level. Since the wave velocity is roughly proportional to the ice velocity, these are approximately places where the ice velocity increases in the down-glacier direction, that is, places of extending flow. The behaviour at places where the wave velocity decreases down-glacier (dc_0/dx negative), that is, approximately, places of compressive flow, is entirely different. Here it can be shown that the glacier surface begins to depart exponentially from its original level. Once lowered from its equilibrium level, it spontaneously falls still

further. In other words, whereas a region of increasing wave velocity is stable, a region of decreasing wave velocity is inherently unstable.

Let us emphasise that this rather unexpected result follows from only two assumptions: incompressibility, and that the discharge is approximately a function of the thickness. There are no specific assumptions about the long profile of the glacier, the roughness of its bed, the flow law of ice, the law of sliding, or about the distribution of accumulation and ablation; although, of course, it is factors such as these which decide whether any particular region will have an increasing or decreasing wave velocity.

We may summarize the position by saying, roughly, that regions of extending flow are stable and regions of compressive flow are unstable.

3. A PARTICULAR MODEL

The consequences of the behaviour just described are most clearly brought out by considering now a particular glacier in which the wave velocity for the steady state, c_0 , is as shown in Fig. 1. x is again the distance down the glacier. In the upper part of the glacier c_0 increases at a uniform rate and in the lower part, below P , c_0 decreases at an equal rate. Since c_0 is roughly 4 times the ice velocity, the graph may also be read roughly as a hypothetical distribution of ice velocity down the glacier. Suppose the glacier starts in a steady state, in equilibrium with the prevailing rate of

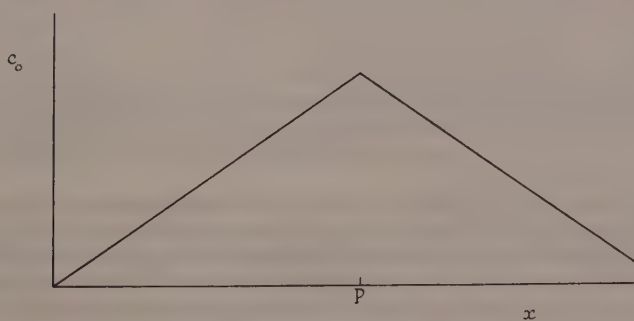


Fig. 1 — Particular model used for detailed calculation, showing the variation of wave velocity for the steady state, c_0 , with the distance x down the glacier.

accumulation and ablation, and suppose that by some means (say by a sudden but temporary, worsening of the climate) a small, uniform, extra thickness of ice (h_1^*) is added everywhere. Figure 2 shows the subsequent behaviour; the extra thickness h_1 , over and above the equilibrium thickness, is plotted against the distance down the glacier. In the upper part the surface begins to fall exponentially to its original position, as one might expect, while in the lower part the surface begins to rise exponentially. As soon as this occurs, there is a region around P where a step would develop, were it not for the fact that such an incipient step immediately begins to move down the glacier as a kinematic wave. The calculated result is that, as shown in the figure, at any place below P the surface continues to rise exponentially until the kinematic wave from P reaches it; it then subsides exponentially. Thus the glacier surface finally returns to its original level; in this sense, the glacier as a whole is stable against perturbations. However, the interesting point is that, in a long glacier, the snout end can grow unstably very considerably before the kinematic wave from P reaches it. For

example, if the ice velocity at the snout is one-tenth of the maximum ice velocity in the glacier, the added thickness at the snout will grow to 40 times its initial value before subsiding again. The whole process reminds one of the crack of a whip.

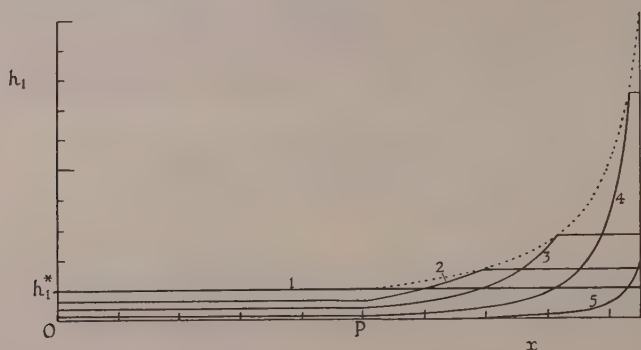


Fig. 2 — The result of a perturbation of the model glacier of Fig. 1. The increase h_1 in thickness is plotted against distance x down the glacier. Initially there is assumed to be a uniform thickening of arbitrary amount (curve 1). Curves 2 to 5 show successive stages in the return to a steady state. For a glacier with a strain-rate of about $+3\%$ per year above P and -3% per year below P (which gives a time-constant of 10 years), curves 1, 2, 3, 4, 5 correspond to times 0, 5, 10, 20, 40 years respectively. The dotted curve shows the progress of the kinematic wave from P .

The time-constant for all these exponential changes (time taken for the change in thickness to fall to 0.37, or rise to 2.7, times its initial value) is $(dc_0/dx)^{-1}$, which is roughly $1/(4r_0)$, where r_0 is the rate of longitudinal strain in the glacier. Thus, if the strain-rate is 10% per year, the time-constant is about 3 years. For a strain-rate of 1% per year, it is about 30 years.

We learn from this example that the lower part of a glacier is in a very delicate balance indeed. By itself it is simply unstable; it is only made stable by the presence of the upper regions, which are continually averting a catastrophe, in one direction or the other, by sending down the requisite kinematic wave to restore equilibrium. We need hardly be surprised any longer at the temperamental behaviour of many glacier snouts, placed as they are in this embarrassing predicament.

It is instructive to consider again the same glacier (Fig. 1), but this time to make a different type of perturbation. We start with the glacier in a steady state, with a certain steady rate of accumulation and ablation, which varies in some unspecified way down the glacier. The rate is then suddenly and permanently changed to a new steady value, in such a way that the rate of accumulation increases, or the rate of ablation decreases, by the same amount all over the glacier. This will mean a sudden permanent lowering of the firn line. The question is: how does the glacier adjust itself to this sudden worsening of the climate? Fig. 3 shows the successive stages. All parts of the glacier begin to thicken, but, whereas the upper part approaches a new steady-state thickness in a stable way, the lower part begins to thicken unstably. This unstable thickening continues at any point until the arrival of the kinematic wave from P . After this the surface continues to rise, but at a decreasing instead of an increasing rate. Eventually a new steady state is achieved, shown by the broken line in Fig. 3. It is noteworthy that the final rise in level of the lower part of the glacier increases markedly towards the snout. For example, if the ice velocity at the snout is one-tenth of the maximum ice velocity in the glacier, the final rise in level will be 79 times greater at the original position of the snout than in the upper part of the glacier. The resulting

advance of the snout down the valley is readily calculated (*). This example shows very clearly the great sensitivity of the snout of a glacier to a climatic change.

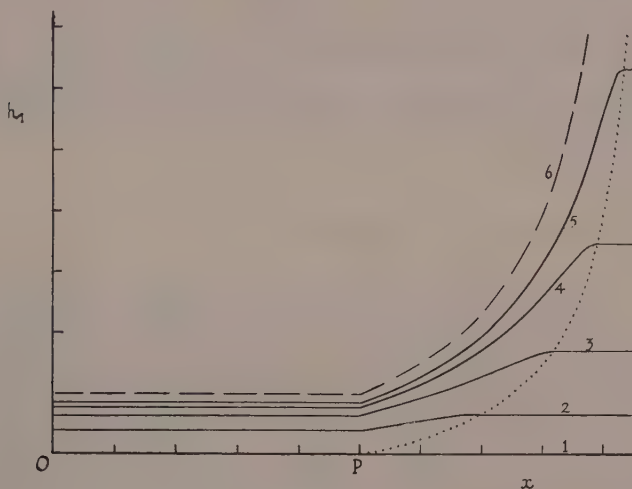


Fig. 3 — The same as Fig. 2, but with different initial conditions. The glacier starts in a steady state (curve 1, i.e. $h_1 = 0$), and there is then a sudden permanent uniform increase in rate of accumulation. Curves 2-6 show the successive stages in the thickening of the glacier. For a glacier with a strain-rate of about $+3\%$ per year above P and -3% per year below P (time constant 10 years), curves 1-6 correspond to times 0, 5, 10, 15, 20, ∞ years respectively. The dotted curve shows the progress of the kinematic wave from P .

It will be seen by a study of Fig. 3 that the thickening at the snout is *not*, in this case, caused by the arrival of a wave in the usual sense. In fact, the arrival of the wave from P is the signal for the return of stability. Further consideration also shows that the thickening at the snout is caused just as much by the decreased ablation below the firm line as by the increased accumulation above it.

4. SUDDEN INCREASE IN RATE OF ACCUMULATION. GENERAL CASE

The example in which the wave velocity c_0 is given by Figure 1 was chosen because it illustrates some of the facts very clearly. However, the theory may equally well be applied to the general case where c_0 varies in an arbitrary way down the length of the glacier. The result of a sudden increase either in the rate of accumulation, or in the level of the surface, is then as follows (unpublished). Regions where dc_0/dx is positive are initially stable, and regions where dc_0/dx is negative are initially unstable—these

(*) If the overall increase in the rate of accumulation (and the decrease in the rate of ablation) is a_1 , it can be shown that the final rise in level of the upper part of the glacier is approximately $a_1/(4r_0)$. Thus, if the rate of accumulation increases by 1 m of ice per year and the strain-rate r_0 is 10% per year, the surface of the upper part of the glacier rises by 2.5 m. If r_0 were only 1% per year the rise would be 25 m.

The eventual increase in length of the glacier expressed as a fraction of the original length turns out to be simply a_1/A_0 , where A_0 is the rate of ablation at the snout. Thus, if a_1 is 0.5 m of ice per year and A_0 is 5 m/yr, the total length of the glacier eventually increases by 10%.

are roughly the regions of extending and compressive flow respectively. However, their stability or instability is not necessarily permanent. An unstable place becomes stable when the kinematic wave arrives which started at the original up-stream end of the unstable region (where $dc_0/dx = 0$). Similarly, a stable place becomes unstable when the kinematic wave arrives which started at the original up-stream end of the stable region. One may thus think of regions of stability and instability which are propagated down the glacier at the kinematic wave velocity. In the examples of Fig. 2 and 3 the forward end of the stable region moves down until the stable region eventually covers the whole glacier.

5. GENERAL THEORY INCLUDING DIFFUSION

A refinement of the theory is to recognize that the discharge at a given point on the glacier will not be determined solely by the prevailing thickness of the ice, but also partly by the slope of the upper surface. The effect of this is to modify the previous results to some extent. It has the general effect of broadening any surface disturbances. Mathematically, the broadening of a disturbance due to this effect obeys the diffusion equation, and we speak of diffusion of the waves, just as one speaks of the diffusion of a temperature hump in a solid, or of a concentration hump in a solution. Diffusion thus sets a limit to the sharpness of a kinematic wave; for example, in the solutions of Figures 2 and 3, diffusion would blunt the sharp corners shown on the curves.

No explicit analytical solutions taking diffusion into account have so far been obtained; but the relevant equations, including diffusion, for the development of any disturbances produced by climatic change (change of rate of accumulation or ablation) can always be solved by numerical integration. The climatic change does not have to be assumed sudden; the rate of accumulation or ablation may vary in any way one cares to specify, both with respect to time and with respect to distance down the glacier. In this sense the theory is entirely general.

6. RESPONSE TO ANY CHANGE IN RATE OF ACCUMULATION WHICH IS UNIFORM OVER THE GLACIER. FOURIER ANALYSIS

Suppose we start with some arbitrary distribution of rate of accumulation and ablation down the length of the glacier. It is then a useful simplification to examine the case where the *changes* of rate of accumulation, produced either by winter and summer, or by year-to-year fluctuations, or by longer-term climatic change, are uniform down the glacier. Thus the rate of accumulation or ablation always changes by the same amount at all points down the glacier.

The change in the rate of accumulation can be resolved into its Fourier (harmonic) components; there will be a strong component with period equal to one year, and also various other components of shorter and longer period. We can examine the response of the glacier to any specified Fourier component, and we then obtain the complete response by simply adding together the separate responses to the various components. The mathematical problem thus reduces to the analysis of the response to a single Fourier component; that is, to a cyclic change of rate of accumulation of arbitrary period. Note that we are not assuming the changes to be cyclic, but are simply analysing them into their cyclic components for convenience.

The result of this analysis is as follows. A glacier has a characteristic response time, as we have seen, which is roughly inversely proportional to the rate of longitudinal strain. This varies from point to point down the glacier, but for typical strain-

rates of between 1 and 10% per year the response time is between about 3 and 30 years. The changes in rate of accumulation caused by the alternations of winter and summer are thus of comparatively high frequency. It then follows that the oscillations in the thickness of the glacier which result from the seasons lag almost 90° behind the oscillations of the rate of accumulation: if the rate of accumulation is greatest in the winter, the thickness is greatest in the spring.

On the other hand, *climatic* changes will have periods which may be comparable with the range of response times of the glacier, or much longer (one should strictly compare the period with 2π times the response time). It turns out that for these changes we have to distinguish between the upper and the lower parts of the glacier. The distinction is most clearly brought out by considering again the model of Fig. 1 (uniform extending flow and uniform compressive flow). In the upper part, the changes of thickness due to climatic changes tend to come into phase with the rate of accumulation—in contrast with the 90° phase lag for winter-summer changes. This is just as one would expect—the glacier is thickest when the climate is worst.

In the lower part, however, the response to climatic changes is more complicated. In the simple model of Fig. 1 the response may be thought of as made up of two separate effects. There is first a *direct* response; and, because of the inherent instability of a compression region, the thickness oscillations due to this effect tend to be in antiphase with the rate of accumulation. There is also a *delayed* response which is propagated down the glacier as a travelling wave-form. The amplitude of the wave changes as it travels; the inherent instability of the region tends to make it grow, but diffusion tends to diminish it. The effect observed at any point on the glacier, or at the snout, will depend on the combination (interference) between the direct and the delayed responses. A wide variety of resultant responses is possible, depending on the distance of travel of the wave, its period, the diffusion coefficient, and so on—but the response in any given case is nevertheless calculable. Thus we begin to understand why it is that the glaciers in nature show such a rich variety of individual responses to climatic variations.

When the same analysis is applied to the Antarctic or the Greenland ice sheets, the response time is found to be roughly 5000 years. Thus, for climatic cycles with periods up to, say, 1000 years, the maximum thickness will occur about a quarter of a period after the worst climate. But for much longer-period changes the time lag approaches 5000 years (the phase lag approaches zero). As regards the amplitude, a sustained increase of 10% in the rate of accumulation over Antarctica would cause about a 60 m rise of the surface.

7. A VALLEY OF GENERAL CROSS-SECTION. OBSERVATIONAL TEST OF THE THEORY

The extension of the analysis to the case where the glacier is in a valley of general cross-section and varying width is straightforward; the discharge at any given cross-section is assumed to be a function of the level and the slope of the upper surface. The main effect is to modify slightly the expressions for the velocity and diffusion coefficient of the kinematic waves. The effect of tributaries is readily included.

The basic variables in the general case, and the quantities which should be measured for a thorough test of the theory, are as follows. They are all functions of distance x down the glacier, and of time:—

Q , the discharge (the volume of ice per unit time passing through a transverse section of the glacier);

B , the breadth of the glacier at the surface;

h , the height of the surface above an arbitrary datum level (the stage);

a , the rate of accumulation or ablation averaged along a transverse line;

α , the slope of the upper surface.

B , h , a , and α may all be measured by standard methods. It would be best to estimate them by making measurements along the central long profile and along a number of transverse profiles. Failing pipe experiments, Q will have to be estimated by measuring the depth and surface velocity along the transverse and longitudinal profiles. The measurements could profitably be done at yearly intervals; as a simplification it would be possible to choose a glacier of approximately uniform width and to confine all measurements to the centre line.

The above measurements would be used to deduce the discharge Q as a function of the stage h and the slope α at each section. This would give the theoretical values of c at each section of the velocity c and the diffusion coefficient D for kinematic waves, for which the equations are:

$$c = \frac{1}{B} \frac{\partial Q}{\partial h}, \quad D = \frac{1}{B} \frac{\partial Q}{\partial \alpha}.$$

It would then be possible to calculate theoretically the changes in the glacier surface that should result from the observed changes in rate of accumulation, and these predicted changes in profile would be compared with the changes observed. If the theory were shown to be valid, c and D would then be known for this glacier for all time, and it would only be necessary to measure the rate of accumulation to be able to predict its entire future behaviour.

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ABOUT THE STATE OF SOME OF THE TIEN SHAN GLACIERS DURING THE PERIOD OF THE INTERNATIONAL GEOPHYSICAL YEAR (IGY)

R.D. ZABIROV

SUMMARY

The Alpine physico-geographical station of Tien Shan, the Academy of Sciences of the Kirghiz S.S.R., began since 1955 to carry out observations on the state of the Tien Shan glaciers. In connection with the I.G.Y., this work obtained a considerably greater scope. Field survey investigations on glaciers, with elaboration of certificates, were supplemented by stationary observations and by reiterated instrumental surveys. In survey investigations were used air-photographs of past years.

In the course of the work, under the I.G.Y. program, were visited and studied 32 glaciers, 31 of which were surveyed with aid of a phototheodolite, and eleven glaciers — surveyed twice.

On key plots were studied (if possible) in detail the entire glaciation, in others — only spot checked separate glaciers.

The studied glaciers are situated in different parts of Tien Shan, on slopes of various mountain ranges, at considerable distances from one another and under diverse conditions of physical geography.

The development of glaciers in these regions proceeds in different ways, though the general background of glaciation degradation remains the most characteristic one for all the Tien Shan regions.

Upon the range Kok-Shaal is situated the Aksaisky glaciation knot, with the area of ice and firn of about 640 km². In relation to orographic reservedness and particular features of relief and climate here are developed great valley glaciers with vast feeding regions, with short high lying tongues and a very high position of the snow line. A characteristic is given of the present-day of two glaciers, typical for this region.

In the basin of the river Kuilu in Central Tien Shan, with a sharply dismembered relief type, are developed typical valley glaciers with altitudes of ends from 3500 to 3700 m and altitudes of the snow-line from 4000 to 4100 m. A characteristic is given of one valley glacier, being in a deep phase of degradation.

At the sources of the river Sarydjaz, in the region adjoining the powerful glaciation knot Khan-Tengri, are considered two glaciers in a stationary state, i.e. that have come to a conformity with the present-day climatic conditions, and two other glaciers, that have undergone a catastrophic advance in the course of the past ten years.

The Academy of Sciences of the Kirghiz S.S.R. has organized observations on the state of glaciers of the Tien Shan since 1955. The Tien Shan physical-geographical Alpine station, which has been carrying out these observations from 1956, was included in the number of those stations which have been carrying out these observations according to the program of the I.G.Y.

In connection with this, the observations over the state of glaciers were expanded. In addition to field itinerary investigations of these glaciers and compiling of passports, station observations and repeated instrumental surveys were made. This work has given us the possibility of collecting important material concerning a great area of the Tien Shan. More than 130 glaciers have been visited and investigated and 31 of them have been photographed by means of a phototheodolite. The groups of the investigated glaciers are considerably remote from each other and located on different slopes of mountain ranges and in different physical-geographical conditions. In connection with this, we can note that though the development of glaciers of these areas proceeds differently, the general background of degradation of glaciation is the most characteristic feature for all the areas of Tien Shan, as well as, for the most part, of the highlands of the terrestrial globe in general.

Recently, in this general background of degradation of glaciation, we notice cases of decreasing rates of recession of separate glaciers. Already mentioned are the cases of unusually rapid advance of some glaciers in Karakorum and Pamirs. We have information about the slowing of the process of recession of a number of glaciers of French and Italian Alps, of the Pamirs and the Tien Shan. But by these fragmentary facts we can't judge the dynamics of glaciation in general. We can very often encounter cases when one and the same mountain system and even one and the same valley has retreating and advancing glaciers. Therefore, we can't solve the problem about the general direction of the process of glaciation without profound analysis of natural conditions.

We should hope that the common work of the scientists of the whole world resulting in the important material which has been collected during the period of I.G.Y. will help us to solve this problem.

In this report on an example of characterization of some glaciers of the Tien Shan, we want to show how varying may be the stages of development of separate glaciers in one and the same mountain system and at one and the same time.

1. TWO GLACIERS ON THE MOUNTAIN-RANGE KOCK-SHAAL IN INTERNAL TIEN SHAN

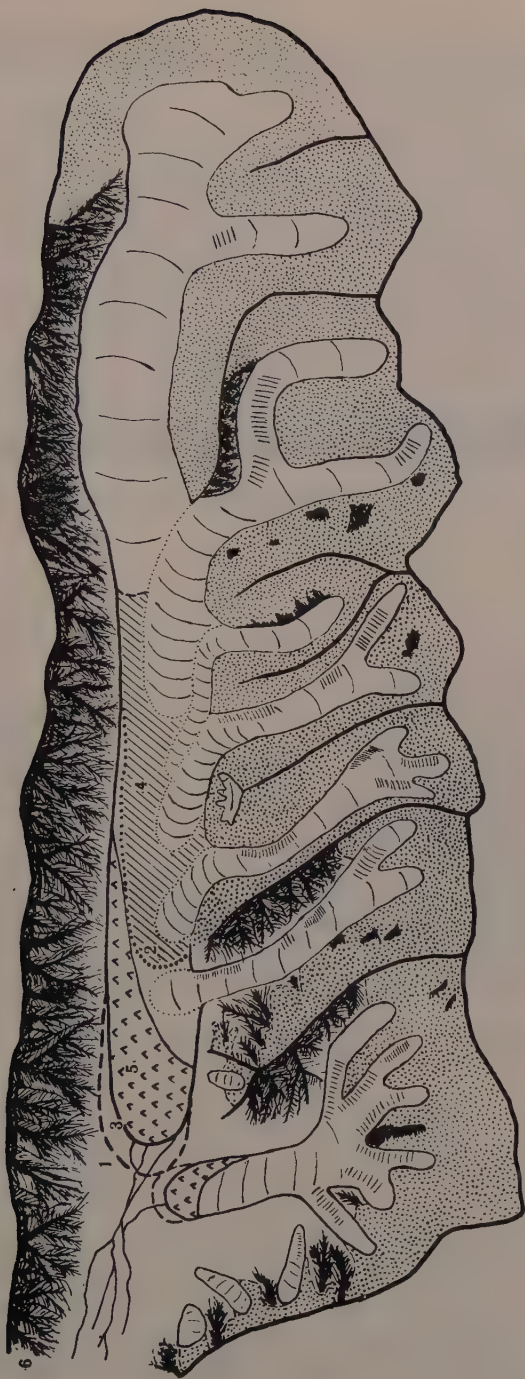
The large Aksaiski centre of glaciation is located in the southern part of the Tien Shan on the Kock-Shaal mountain-range. The height of the mountain is sufficient for the development of large glaciers in this area, even when the precipitation is very small. The glaciers and snow-fields occupy more than 600 sq km. Large Alpine glaciers are located here:

1. Ototash — 13.2 km in length
2. Itala — 11 km in length
3. Komarova — 9.2 km in length

In 1956 and in 1957, the ends of some glaciers of this area were photographed by means of a phototheodolite and described in detail in order to clear up the question about their state during the period of the I.G.Y.

The first two glaciers—Ototash and Itala—are the most interesting. The first one is located in the area of an ancient surface of denudation and the second one in the area of erosion. Therefore, they are quite different in the height of their ends and the slope of their surfaces, etc... The Ototash glacier is located on the northern slope of the mountain-range Kock-Shaal and is the largest glacier of this area—it is 13.2 km in length and the size of its area is 23 sq km. Along the end of the glacier there is a fan-like swell of morainic hills, scattered in disorder, together with lakes. The width of the swell is 25 metres and the height of the hills is 20-25 metres. The outside edge of the morainic swell is sharply outlined and gives place to broad, pebbly flood-lands. The end of the glacier is located at the height of about 400 metres. The glacier has an ablong form, but its feeding area is not developed well. All the way along the ridges of the glacier, it is completely covered with snow, with the exception of the lower section of the slope of the valley of 3 km in length. The primary mass of snow moves from these slopes, therefore, we can state, that this is a special mountain-valley type glacier with double-sided slope feeding. Such type of feeding is formed when the correlation between the snowline and the conditions of the relief is favourable (fig. 1).

Owing to comparatively gentle slopes, snow does not slide as avalanches, but is accumulated upon the slopes as compact cover and gradually streams down into the valley-bottom. In most cases, such accumulations of snow on the slopes are joined with neighbouring glaciers by means of water-sheds.



g. 1 — The glacier of Mushketov and Bezimani. 1. The position of the Ridge of the glacier in a period of a maximum advance until 1943. 2. The position of the Ridge of the glacier in 1943. 3. The position of the Ridge of the glacier in 1959. 4. A zone of «dead» ice in 1943. 5. A pressure swell from «dead» ice and moraines. 6. Rocks. 7. Snow-covered slopes and snow-fields.

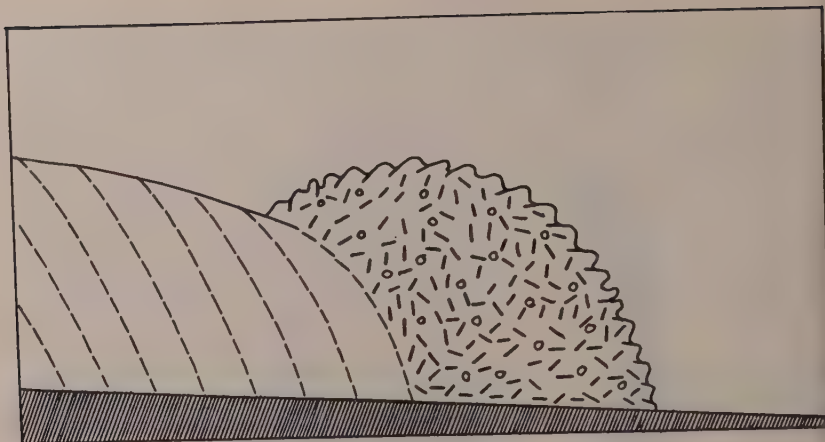


Fig. 2 — A profile across the pressure swell.

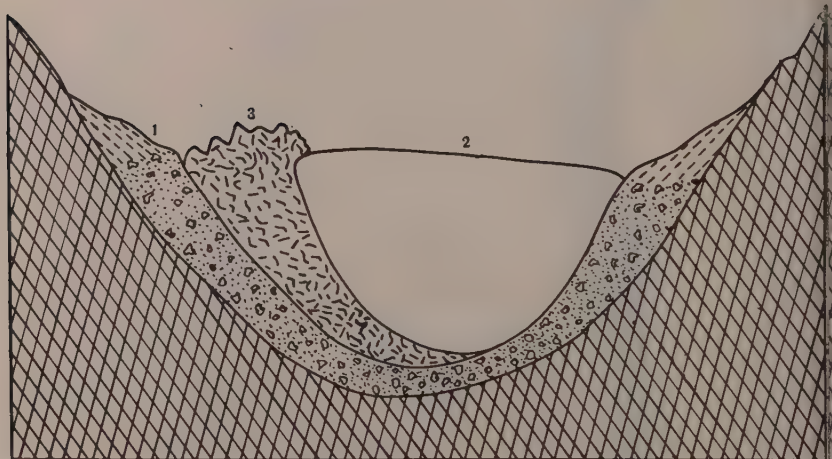


Fig. 3 — A profile across the end part of the glaciers Bezimani. 1. The old sidemore. 2. The glacier. 3. The side pressure swell from «dead» ice.

It is quite probable that glaciers with such a type of feeding have special conditions of movement since they are subjected to constant powerful double-sided pressures. This feeding results in absence of side and middle moraines in the glacier of Ototash. Its end morainic swell consists of an ablation moraine, which covers the surface of the terminus of the glacier even in present times.

The thickest part of the glacier, which is 3 km from the terminus, was calculated by the difference between the slope of the bed and the surface of the glacier. It achieves 80 metres and near the terminus is 30-40 metres. The surface of the glacier is smooth and has no crevasses, splits, hills and gaps. The end of the glacier forms a comparatively steep forehead, along which numerous streams are flowing down. Even in the zone of 3 km the snow on the surface of the glacier melts away only for a very short time (1.5-2) months.

In a period of preparation for the I. G. Y., the end of the glacier has been photographed by means of phototheodolite camera from a baseline, the length of which is 102 m.

In 1957 we made a second survey of the glacier for the estimation of those changes which had taken place at the end of the glacier. The comparison of stereopairs of the repeated surveys, showed us that the glacier was slowly retreating and had decreased in thickness. It appeared to have retreated 3-4 metres a year and to decrease in thickness 0.8-0.2 m. The surface rates of movement also appeared to be very insignificant—9 metres a year. The left part of the glacier, which is the least mobile, became thinner than other parts of it. The investigation of the end moraines showed that before this time, the glacier of Ototash was also retreating very slowly and, therefore, this glacier is believed to be a stable glacier. Its reaction towards the changes of climatic conditions is very weak. The stability of the glacier is due to its development in very high zones, where the amount of precipitation is small and the temperatures are low.

The second glacier of the described area, Itala, is located in a deep narrow valley. In former times its tongue ended 300-400 metres lower than the glacier of Ototash. This is connected with the fact that this area is subjected to strong regressive erosion and its valleys deeply cut into the surface of ancient denudation and the glaciers have slid down the bottoms of steep erosional valleys. Here we can observe a very interesting picture of combination of narrow erosion gorges, flat steps and watersheds with snow-caps.

Like the above described glacier, the glacier Itala has an oblong form, but in contrast with the former one, the end of it seems to enter a narrow gorge; not long ago it was 3600 m lower. At present, the glacier has retreated by 1700 metres. Its end is located at a height of about 3800 metres. The thick frontal moraine of the glacier is squeezed in a narrow erosional canyon, along the sides of the valley it stretches like a coastal moraine and further continues like a side moraine.

There is a broad sandy field between the edge of the glacier and the frontal moraine. Near the sandy field the end of the glacier wedges out, and higher the body of the glacier is divided into separate blocks and represents a disorderly field of "dead" ice. The glacier has been retreating for a very long period of time. 30 years ago when the Soviet glaciologist Palgov N. N. visited this glacier it also had been shrinking. By 1957, in contrast to the Ototash glacier, it had retreated almost 2 km and its thickness had decreased 40-45 metres. It is quite probable that this recession will continue until its end nearly approaches the step of the ancient denudation surface, which is not subjected to erosion.

2. A CATASTROPHICALLY RAPID RETREATING GLACIER IN TIEN SHAN

In the basin of Kuilu river in Central Tien Shan, where the relief is sharply dismembered, typical valley glaciers are developed. The heights of their ends achieve 3300-3700 metres, and the height of their snow-line—4000-4100 metres. The glaciers and snow-fields occupy 29% of the basin area. Almost all the glaciers of this valley are in a stage of degradation; small ones are shrinking in a lesser degree, large ones in a greater degree.

One of the most rapidly retreating glaciers is the glacier of Karakoltor. In comparatively recent times it was a typical valley glacier with some right and left outlet glaciers. The first left tributary was a typical Alpine glacier 5 km in length and the others moved out of cirques.

The watersheds surrounding the glacier Karakoltor have typical Alpine forms of relief: numerous cirques, cutting into the walls of slopes, dismember them into a labyrinth of sharp peaks with steep and sheer walls. In comparatively recent times,

the glacier was 10.5 km in length. It was believed to range from 115-120 metres in thickness in its lower zone from 3-4 km. We have reliable data about this glacier only from 1943. But from many facts, it is evident that the recession of the glacier began much earlier. The sections of the valley-bottom and of the slopes are distinctly outlined. They are distinguished by the whitish tinge of fresh denuded moraines not yet having grass upon them and not covered with desert sunburn. These features are characteristic for Tien Shan. They help us to determine the extent of the recession of the glacier and its thickness in the retreated section. The above noted features help us in stating that by 1943 the glacier Karakoltor had retreated by 1250 metres in this section; and by almost 100 metres in thickness.

The thickness of the remaining part of the glacier greatly decreased; for a distance of 2-25 km from the end of the tongue it thinned only by 10-20 metres. Evidently, this part of the glacier presented "dead ice" which had no outlet glaciers from the feeding area. At the same time the three right and the two left outlet glaciers have separated and turned into independent glaciers. All this resulted in the fact that from 1943 the glacier has begun to retreat very rapidly. As a matter of fact this was not the recession of non "alive" edge of the glacier, but it was rapid melting of a comparatively thin sheet of "dead" ice. During the last 13 years it has retreated, in average, 90 metres a year and has shrunk 1200 metres in length. The next three years, the glacier has additional 400-450 metres. Thus, by 1959 the glacier had shrunk 2900 metres in length. The ice in the retreated part had 120 metres in a maximum thickness. It is very likely that this value is characteristic for the remaining part of the glacier tongue, except for the 1,5-2 km lower zone.

Judging by the character of the lower part of the tongue we can suppose that it will shrink some 1200-1300 metres more if the climatic conditions remain without change.

3. CATASTROPHICALLY RAPID ADVANCING GLACIERS ON TIEN SHAN

Due to the itinerary investigations of the glaciers of central Tien Shan, according to the program of the I.G.Y. in summer of 1959, we found two glaciers subjected to exceedingly rapid advance. The larger one goes by the name of a well-known Russian geologist and explorer of Middle Asia, Mushketov, the second one is a comparatively small valley glacier and has no name.

Both glaciers are located on the Northern slopes of the Sarijas ridge and are the sources of the river of the same name. The Mushketov glacier stretches along the foot of the ridge in latitudinal direction from the East to the West. In this part of the slope the Sarijas ridge is very high and the layer of snow is extremely thick. Long and narrow streams of ice flow down the valley floor, then coalesce and form the glacier Mushketov. There are no snow-fields and glaciers upon the right side of the valley which is rocky (fig. 1) and denuded. The glacier has its feeding controlled by its outlet glaciers. Therefore, any change of these outlet glaciers affects the main glacier. In spite of its large dimensions (it is more than 15.5 metres in length and about 13 m in breadth) the glacier of Mushketov will cease its existence very quickly if the outlet glaciers are separated. In a period of its maximum development, the glacier was 18 km in length, and more than 100 metres in thickness (at a distance of one km from the end).

Later, the glacier gradually began to retreat. The thickness of ice gradually began to decrease and by 1943, the glacier had shrunk by 3 km and the remaining part of the tongue presented an area of "dead" ice for a distance of 6.4 km. The five lower outlet glaciers moved into the area of "dead" ice in form of streams, resembling "lion's paws".

Taking all these facts into consideration, concerning the state of the glacier, we might suppose that these areas of "dead" ice would disappear in a very short space of time, and the outlet glaciers would be separated into independent glaciers. This, however, has not happened. Before the sections of "dead" ice melted, the glacier began to advance very rapidly, crushing and breaking everything on its way. By 1959, the glacier had advanced by 2458 metres. All the "dead" section of the glacier has turned into a heap of debris, girdling the end part of the glacier in the form of a swell. Blocks of ice were piled up one upon the other, some of them resembled sharp peaks, blades, wedges and steeples; there were gaps, cracks, grottoes and passages among them. There were black and dirty blocks of ice as well as clean ones. These blocks were constantly breaking into small fragments and falling down, disappearing in gaps.

The sheets in all these blocks of ice are in a mess: some of them in vertical and horizontal position, others under various angles of incidence. Along the right edge of the glacier, the same picture of disorderly heaped up blocks of ice from 10 to 15 metres in height has been observed. The same relief has been observed for a distance of some km.

The swell from the blocks of "dead" ice is distinctly separated from the "alive" part of the advanced glacier. The latter is distinguished as one stream of ice with alternating transverse crevasses and belts of moraines. The formation of a swell from ice shows that the "dead" end parts characteristic of retreating glaciers have no dynamic ice-exchanging connection with the upper zone. Owing to this "dead" ice and loose depositions under the pressure, the "alive" part of the glacier is transported down along the edge of the glacier.

During the last 16 years, the glacier of Mushketov has been advancing with an average speed of 154 metres a year and for the same period of time its area has increased more than 4 sq km. The advance of some glaciers of Karakorum and Pamirs is supposed to have taken place not in a period of these 16 years but only recently or at the end of it. We can judge from it as an example of some glaciers of Karakorum and Pamirs.

The second advancing glacier is located on the northern slope of the mountain-range Sarijas, near the end of the glacier Mushketov. It is 5 km in length. This glacier is also considered to be a complex one, because it is formed by the coalescence of six small ice streams.

In 1943 it was a retreating glacier with a flattened tongue, but by 1959 it had advanced 540 metres and entered the valley of the Mushketov glacier. The thickness of ice had increased. There was observed the same picture of disorderly heaped up blocks of "dead" ice at the end part of the tongue and at the right side of the glacier, and a pressure swell from the blocks of ice above the surface of the glacier. It is 10 metres thick.

We are not prone to attribute the catastrophical advance of the above described glaciers to essential changes of climatic conditions. On the contrary, it is connected with slight local changes of conditions of feeding of outlet glaciers which has produced a sharp change in conditions of the development of a main glacier. As stated above, the separation of the outlet glaciers from the glacier of Mushketov might have led to almost complete disappearance of it. On the contrary, minimum increase of the outlet glaciers may promote the advance of the glacier, because this will lead to a disturbance of the accumulation-ablation balance.

We observe the same picture on the glacier of Mushketov. The breadth of the main stem of the Mushketov glacier is three times less than the total breadth of all the outlet glaciers. It means that the unit of volume of ice in the main stem of glacier has three times less area of melting. Hence, it follows that in comparison with outlet glaciers the accumulating ice is three times higher. This results in the formation of excess reserves of ice which promote the increase of thickness and the advance of the glacier. This phenomenon takes place until the equilibrium is reached.

The Tien Shan high-mountain physico-geographical station of the Academy of Sciences of the Kirghiz S.S.R.

EVOLUTION OF SOME TIEN SHAN GLACIERS DURING THE LAST QUARTER OF THE CENTURY

L.G. BONDAREV

SUMMARY

1. Tien-Shan, as well as the Pamirs, is the greatest region of present-day mountain glaciation in the limits of U.S.S.R. The massif Akshiyriak is the third centre of glaciation on the Tien-Shan as to size after the region of the peak Pobieda (Victory) and Khan-Tengry. Approximately one half of the massif's territory is covered by glaciers and firn fields.

2. During the period of the second International Polar Year (1932-1933), some of the glaciers of this region were described by the Naryn-Khan-Tengry expedition.

In 1957-59 almost all these glaciers were visited by survey detachments of the Tien-Shan alpine physico-geographical station of the Academy of Sciences of the Kirghiz S.S.R. A comparison of newest observations with data, obtained during the second I.P.Y., and also with air photos of 1943, allows to make conclusions about the evolution of Akshiyriak's glaciers during the last quarter of the century.

3. The majority of glaciers has been reduced. Diminution in length of certain glaciers is calculated by hundreds of meters and sometimes it exceeds 1 km.

4. The diminution of the glaciers' dimensions is accompanied by their flattening. This is testified by the fresh ridges of lateral moraines and subsidence terraces. The tempo of flattening of different glaciers is equal to 1.5-2 m per year.

5. Glaciation undergoes also modification the quality order. Noted is a number of cases, when glaciers desegregate into smaller ones. Some glaciers are found to be at present in a state of desagregation. There occurs also an isolation (setting apart) of lateral tributaries of valley glaciers.

6. Degradation of big glaciers is accompanied by an appearance at their terminal parts of considerable particles of dead ice. The latter circumstance will stipulate the continued reduction of linear dimensions in a series of glaciers, even under a « bettering» of climatic conditions.

7. On the general background of the glaciers' regress, seems paradoxical the transgression of the second in size glacier of Akshiyriak — of the Northern Karasai glacier, which took place between the years 1946-49 and 1955-1956. The glacier moved forward by nearly one km. Progress of the Northern Karasai is explained by non-climatic causes. The main flow was on the eve of being transformed into dead ice and moved very slowly. This caused a saddling in the estuaries of lateral tributaries, whose ice was no more involved into the main flow's movement. In result, there took place an ice accumulation in the tributaries' mouth, a rise in their thickness and a swelling all this lead afterwards to a short-time progress of the glacier.

8. It is noted, that degradation involves principally big glaciers, whose tongues descend lower. Small hanging and kar glaciers, those of hanging valleys, have retained in many cases a stationary state or have changed insignificantly.

9. At present, is observed an augmentation of firn reserves in feeding regions of many glaciers. This is why a lagging of the regress tempo can be awaited in the near future; this lagging will be substituted afterwards by a transgression of some of the glaciers.

The Tien Shan as well as the Pamirs is the largest region of contemporary mountain glaciation on the territory of the U.S.S.R. The area of the contemporary glaciation on the Tien Shan is 8727 km² which is nearly 4.5 times as much as the glaciation area of the Alps.

The Akshirak mountain system, consisting of parallel ridges directed to the South-West, is third in size of the contemporary glaciation centres in the Tien Shan, after the Khan-Tengri region and «Aksai group». Nearly half of its territory (43%) is covered with glaciers and neve fields. The area of the contemporary glaciation is considered to be 439.5 km².

During the second International Polar Year (1932-1933) some of the glaciers of this district were described by the Naryn-Khantengri expedition and the terminal of some of the glacier were mapped with the help of theodolite.

The results of this research are described in the works of professor S. V. Kalesnik and other members of the expedition, published in 1935.

In 1957-1959, nearly all the glaciers described during the second International Polar Year were visited by special detachments of the Tien Shan Alpine physic geographical station of the Academy of Sciences of the Kirghiz Republic. The comparison of the new observations with the data of the second International Polar Year as well as with the aerophotography of 1943, enables us to come to a conclusion with regard to the evolution of some of the Akshiral glaciers during the last quarter of a century.

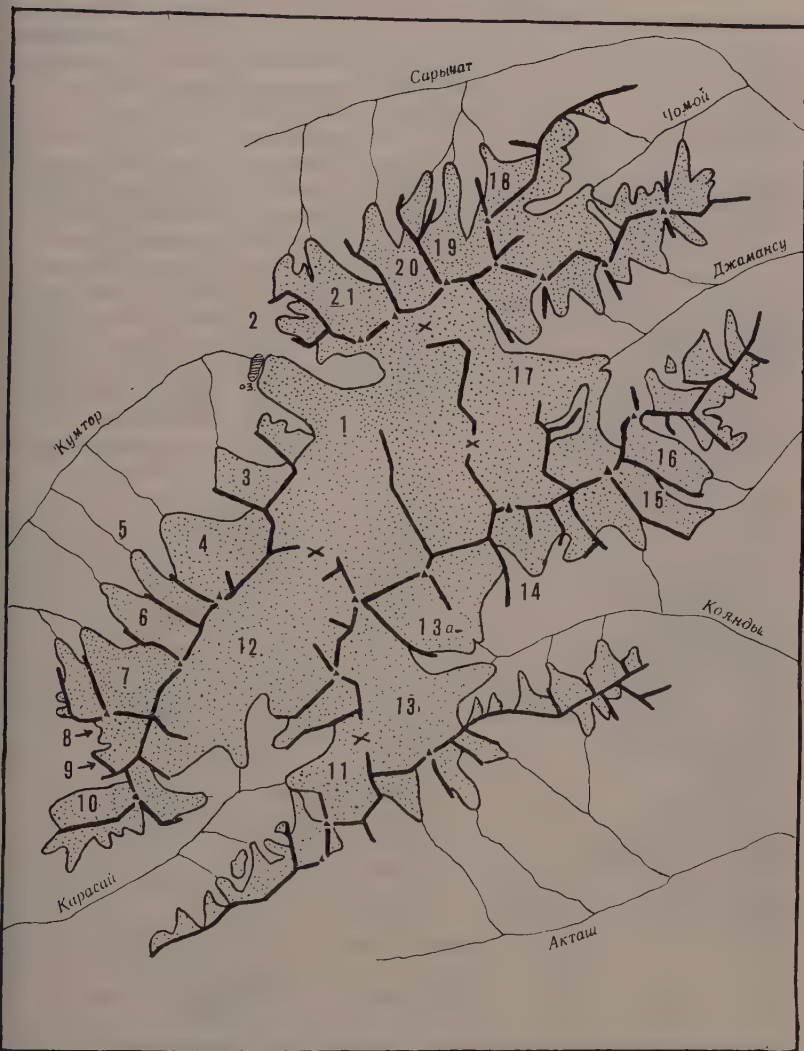


Fig. 1

The members of the second International Polar Year discovered that the Akshirak-glaciation was regressing. Many glaciers continued to shrink after 1932-1933. The data of the lessening of length and of area of 21 (twenty-one) glaciers, the evolution of which we had already studied from 1932-1933, is given in the following table:

Evolution of some Akshirak glaciers from 1932-1933 to 1957.

Names of glaciers	Contem- porary area in km ²	Absolute height of lower glacier terminal	Distance of withdraw- ing of glacier terminal in metres		Area gla- ciers are shortened by in km ²
			1932-1943	1943-1957	
1. Petrov Glacier	70,6	3730	?	400	0,79
2. Double (Dvoynoi) left branch	2,6	3840	70		0,13
right branch		3880	—		
3. Bald (Lysyi)	4,4	3790	Stationary	60-70	0,07
4. Davydov glacier	12,1	3820	290	380	0,51
5. Sarytor — 2	3,0	3820	75	25	0,04
6. Bordu — Northern	5,8	3840	500	+ 260	0,09
7. Bordu — Southern	9,2	3810	120-130	Stationary	0,05
8. Large Kazan (Bols- hoi Kasan)	1,5	3910	30	60	0,02
9. Little Kazan (Maly Kazan)	1,2	3930	15-20		< 0,01
10. Akbel	3,9	3945	80-100		0,05
11. Southern Karasai (Edelstein glacier)	17,2	3750	200-300	700	1,15
13. Koyandy and	38,0	3850	180-200	460-520	0,27
13. a Fan-shaped (Vee- roobrazny)			330-530	190-210	0,70
14. «E» Glacier	6,6	3970	Stationary or moving		—
15. Right Kurgatepchi	5,0	3910	170-220	500	0,23
16. Left Kurgatepchi	4,9	3850	Stationary		—
17. Jamansu	36,4	3545	1050		0,68
18. Right Oroisu	3,6	3970	1090-1100	Stationary	0,69
19. Left Oroisu	8,6	3960	975	150-200	0,51
20. Nameless (Bezymy- anyi)	5,2	3920	800-900	150-200	0,34
21. Sarytor — 3	11,3	3820	400 (main glacier) 780 (former lower tributary)		1,54
Sum total	297,5		Sum total		6,22

The names of the glaciers in the table are the same as those described in the documents of the second International Polar Year.

The lessening of the glacier's lengths is accompanied by their flattening. This can be proved by a wide spread occurrence of new ridges of lateral moraines and terrace subsidence.

According to S. M. Myakhkov's data, the thickness of the Davydov glacier decreased by 21 m or by 1.6 m yearly from 1943—till 1957 (on the cross-section of the lower one-third part of the tongue).

During the same period of time the thickness of Southern Karasai decreased by 36 m or by 2.5 m yearly.

The height of the left edge slope of the Bolshoi Kazan glacier was 30-40 m in 1932 and now it is nearly 20 m.

Because of the fact that the middle moraines preserved the ice, sharp narrow rests, up to 20-25 m high, appeared under them when the terminal part of the Jamansu glacier was flattening. Judging from the plan of the terminal part of the glacier made according to the second International Polar Year documents, the axis parts of the middle moraines barely exceeded the glacier surface in 1933.

The shrinkage of the glaciation is accompanied by qualitative changes. There were cases of some glaciers disintegrating into smaller ones.

Thus, for example, in the valley situated to the North of the Petrov glacier, another glacier (the Double) was described in 1932 which consisted of two branches with connected terminals. The right flow of ice was creeping over the left one. By 1943 the division of the glacier into two separate parts became evident. The two tongues were still connected but this connection was passive. At the place where the right and left branches joined there was a deep hollow.

Apparently, a great role in the division of the glacier was played by erosion. The final disintegration took place between 1943 and 1957. In 1957 we already found two independent glaciers here. The terminal of the right glacier was separated from the left one by about 160 m.

The glacier situated to the North of the Koyandy glacier terminal (which is described in the documents of the second International Polar Year under the name of Veerobrazny) was its tributary in 1933. The tributary had separated from the main glacier by 1943. The right side of the terminus of the former tributary was 330 m and the left one 530 m away from the left lateral Koyandy moraine. In 1957 this distance increased to 540 and 720, respectively. In the terminal part of the former tributary, large sections of «dead» ice appeared and now its division into two independent parts is being outlined. In this case we have an example of exceptionally quick qualitative transformations which have led to the glacier disintegration.

The first lower right tributary of the Koyandy, which in 1932 was connected with the main glacier, has also been isolated.

The valley glacier with its tributaries divided up into three independent glaciers in Sarytor—3 valley between 1933-1943.

Some lateral tributaries of the valley glaciers are now in the state of isolation. Such are the first lower left tributary of the Southern Karasai glacier and the first lower right Jamansu tributary. It is interesting to note that, in both cases, the lessening of the length and the thinning of the main branch caused an advance towards the main branch of the tributaries which showed a tendency towards isolation; this can be explained by the decrease in pressure on the side of the main branch.

The degradation of glaciation is accompanied by the appearance of large sections of dead ice in the terminals of the biggest glaciers.

For several kilometers, the tongue of the Petrov glacier, the largest glacier in the Kshiryak, is, evidently, stagnant ice.

The relief of the glacier surface is very complicated. It abounds in many craters, wells, ablation ditches developed out of large crevasses, and deep canyon-like channels

of meltwater flow. These forms of relief testify that the tongue of the glacier is immovable, that the glacier is unable to fill in the negative forms created by ablation. The forms of relief on the Petrov glacier which are characteristic of immovable and slightly movable ice, had already been observed before: the description of 1932 describes craters up to 80-100 m deep.

The tongue of the Northern Karasai, where the surface of the main glacier is covered with innumerable craters thus making a smallpox-like relief for almost 9.5 km, is evidently immovable, too. Dead ice is also plentiful at the terminals of Jamanus, Koyandy, Southern Karasai, Fanshaped glaciers and at the terminal of the main glacier in the Sarytor valley.

Large sections of dead ice are not typical of glacier terminals of smaller dimensions.

Along side with the dead ice which is a continuation of the active part of the glacier there are plenty of isolated blocks of dead ice which can often be found exposed in the banks of lateral and terminal moraines.

The fact that a number of glaciers have considerable areas of dead ice will stipulate the continuing lessening of their long measures even if the climate somewhat «improves».

There are only three cases of advance of ice. The advance of the Akshirak glacier (Northern Karasai), which took place between 1946-1949 and 1955-1956, appears especially paradoxical. The glacier advanced for almost one kilometer and formed a bank of a terminal pressure moraine, composed of well-smoothed sandstone and shingle deposits.

The Northern Karasai is a glacier of the dendrite type which takes in nine valley tributaries on the left and on the right. The advance of the glacier was caused not by climatic changes, but by the complicated phenomena of the glacier's development.

By 1946-1949 great masses of ice in the main current were on the point of turning into dead ice; the terminal part of the tongue, just where two lower right tributaries flow into the main branch, moving exceedingly slowly or even not moving at all.

The main glaciers block the tributaries at the entrance. In our case, the effect of this blocking increased as the movement of the ice of the main current slowed down since the ice of the tributaries was not involved in the movement of the main glacier but was blocking its own flow.

The accumulation of ice began to exceed its discharge; the power of the tributaries increased, their discharging parts spread, swelled and acquired an influx form. A process took place which M. V. Tronov called «selfgrowing» of glaciers. Blocking could also take place in the part of the main flow of ice higher than the terminal that «stopped» moving.

The masses of ice formed by blocking, possessed a great potential energy and much propensity for flowing. At last the accumulation of the quantitative changes resulted in a qualitative leap—the appearance of a lateral flow of ice. The impulse that caused this change was apparently the heavy avalanches in the névé basins of the tributaries or in the area of the feeding of the main glacier and, as a result of this the shoving into the valley of considerable additional masses of ice.

Thus, the advance of the Northern Karasai was going on mainly at the expense of the ice accumulated during the selfgrowing of the tributaries.

The liquidation of the blocks and the immediate involving of great masses of ice in this movement constituted the sudden qualitative leap; the advance of the glacier ought, therefore, to be characterized by abruptness and rapidity. It is quite possible that the advance of the Northern Karasai was in progress for no more than 1-2 years or even for only a few months. At present, the glacier is shrinking again.

The terminal of the glacier in the Northern Bordu valley was, in 1932, about 500 m away from the lower terminal of the left lateral moraine. In 1943 this distance measured by airphotography was about 1000 m and by 1957 it lessened to 740 m. The glacier has advanced but is still 240 m shorter than in 1932.

The terminal of the glacier in 1932 represented a tattered blade, rather flat and gently sloping into the valley, which is typical of a degrading glacier. In 1943, the tongue had the shape of a plain, monolithic blade. We can infer, therefore, that the advance of the glacier could have begun before 1943.

At present, the terminal of the tongue is broad and obtuse, edged with a steep precipice 10-15 m high. The brow of the precipice is beveled at places and at the bottom of the precipice there are blocks of ice which have fallen from above. All this tells of the considerable speed of movement of the upper layers of ice.

The causes that make the glaciers advance are not clear. It is quite possible that the main cause lies in the increase of the area of accumulation as a result of the blocking that took place on the border with the upper part of the Northern Karasai glacier.

The glacier situated in the first lateral valley of the left slope of Koyandy, to the east of the end of the main glacier (in the documents of the second I.P.Y. named «Glacier E») has kept its stationary state or changed but inconsiderably since 1933. Investigation of this glacier in 1959 showed that the terminal part of the tongue was creeping against dirty and crevassed dead ice.

It is worth noticing the following peculiarity of the decreasing glaciation that is taking place now. This decreasing is going on mainly at the expense of the major glaciers. The glaciers of smaller dimensions change in a smaller degree, whereas small and pendant glaciers, with an area of up to 2 m² have kept their stationary positions since 1943, being, at the same time, in exceedingly disadvantageous conditions of the Southern exposition. (We do not have the information regarding the evolution of the minor glaciers, prior to 1943, at our disposal). The glaciers of minor dimensions correspond better to the present climatic conditions.

This can be explained by the fact that big glaciers have not, up to now, discharged excess ice which was accumulated during a more humid epoch. The comparatively fast rate of withdrawal of the major glaciers is connected also with the peculiarity of their degrading—the presence of big areas of immovable dead ice, the thawing of which is not compensated at all by the influx of the substance from the feeding area.

Deviations from the mentioned regularity are not unexpected, they can be explained by the concrete peculiarities of the glacier—by the local orographical conditions, by the presence of the protecting coat of a surface moraine, by the blocking phenomena and by other factors.

Such glaciers as the Jamansu, the Koyandy and the Southern Karasai would have withdrawn still further had there not been a surface moraine covering the terminals of tongues.

A strange, at first sight, advance of the Northern Karasai finds its explanation in the peculiarities of the development of the glacier.

Large (in per cent ratio) areas of ice were lost at the time of disintegration of the Darytor 3 glacier and the Dvoynoy glacier. This can be explained by the fact that the qualitative leap—the process of disintegrating of the glacier—has been going on quickly and irreversibly.

The causes of rather considerable shortenings of comparatively small glaciers in the Northern part of the Akshirak (the glaciers of the Oroisu and of the Bezmyani asins) are not quite clear.

Maybe this is due to the fact that the terminals of the glaciers had, for a long time, been preserved by a moraine and had kept, therefore, a stationary position. The degradation of the glaciers meant the thinning of the tongues, which resulted, afterwards, in their severing from the terminal of the glacier covered with a moraine.

As a result of this, a rapid withdrawal of the weak, flattened tongue took place in the course of a short period of time. Fig. 2 shows the shortening of a clear glacier (left) and a glacier the terminal of which is covered with a moraine.

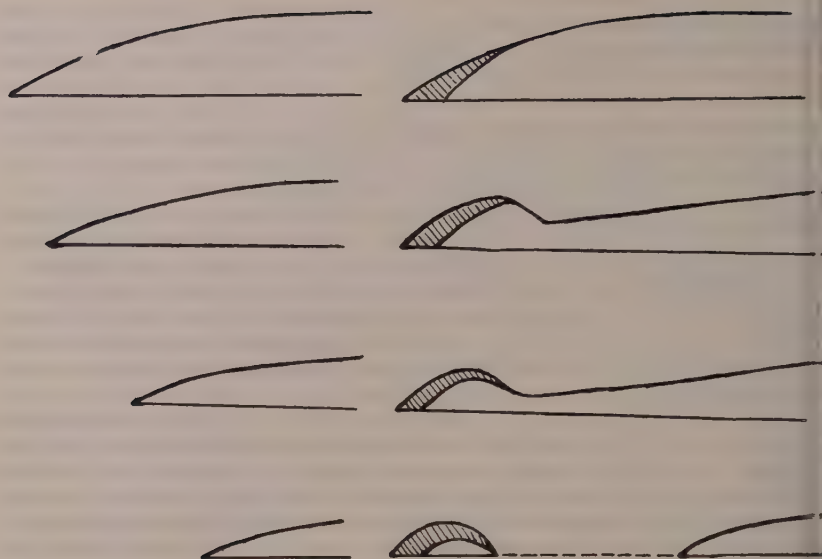


Fig. 2

The withdrawing of the terminal of the first glacier goes on gradually while the lineal shortening of the second goes on by leaps, a short term period of rapid withdrawal being preceded by a gradual thinning of the tongue.

It is quite probable that it might be exactly in this way that the lessening of the above-mentioned three glaciers of the Northern Akshirak was going on and that was just between 1933 and 1943 that a leap in the lineal withdrawing of these glaciers took place. As to Bezymianny glacier and the Oirosu (Right) glacier, we have grounds to suggest a similar course of events, for in 1933 these two glaciers terminated in flattened tongues and were very near to the inner edge of the moraine.

The total area by which the glaciers described by us were shortened equals 16.22 km². Thus, in the course of 25 years, the area of glaciation has decreased by 2.1 per cent, or is lessening by 0.08-0.09 per cent yearly. With a certain error, these figures can be applied to the whole of Akshiryak.

It is necessary to stress, however, that to make a quantitative characteristic of the decrease of glaciation, it is not enough to calculate only the area of the ice that had disappeared, for glaciers lessen not only by withdrawing but also by flattening. The most objective and interesting figures could be received by computation of the bulk of ice by which the glaciation was lessened. But we could not make such a calculation as data on the amount of thinning during the described period are not available.

It is interesting to compare the lessening of the Akshiryak glaciation area with that of the Swiss Alps and the Caucasus. The dynamics of glaciation of these glaciers during the last decades is described in detail. In the Swiss Alps the area of glaciation has decreased by 25 per cent or by 3.5 yearly for half the century.

The area of glaciation in the Caucasus decreased by 8.5 per cent or by 0.15 per cent yearly from 1890 till 1945. Thus, the yearly decrease of the area of glaciation Akshiryak is 8 times less than in the Swiss Alps and almost 2 times less than in the Caucasus.

Our comparison has two shortcomings. In the first place, it is wrong to compare the glaciation of the Caucasus mountain system with the glaciation of only a part of the mountain system restricted by administrative borders (the Swiss Alps) and with the glaciation of a comparatively small mountain group (the Akshiryak) which is a part of a vast mountain system (Tien Shan). And, in the second place, we have compared the dynamic glaciation for different periods of time.

Nevertheless, we may say that the glaciation in Akshiryak has degraded less than in the Caucasus and the Swiss Alps for the last quarter of the century.

At present, in the regions of feeding of many Tien Shan glaciers, one can see the increase of névé supplies. In particular this phenomenon has been marked in the Kulilu basin, a district situated to the East of Akshiryak by R. D. Zabiroy.

Therefore, in the near future, the rate of the glaciers' retreat will apparently slow down and will perhaps be replaced by the advance of some glaciers, especially by those which have been remaining stationary for the last few years.

The tongues of large glaciers, having big areas of dead ice at the terminals, will continue to shrink while their active parts will increase in their thickness and will perhaps creep over the dead ice. The increase in thickness will be stimulated by the pressure of the dead ice.

The Tien Shan physic geographical Station of the Academy of Sciences of the Kirghiz Republic.

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DISTRIBUTION AND VARIATIONS OF GLACIERS IN THE UNITED STATES EXCLUSIVE OF ALASKA

Mark F. MEIER

U. S. Geological Survey, Tacoma, Washington

SUMMARY

A cooperative program of investigations of the distribution of existing glaciers in the United States south of Alaska and the variations of these glaciers was instituted during the International Geophysical Year. Approximately 1,000 glaciers were found to exist; 77 percent of the glacier area occurs in the State of Washington. The total glacierized area is 513 km². Quantitative data on surface rise, advance of terminus, gross accumulation, late summer ablation rate, and measured precipitation were obtained for seven glaciers, and qualitative data were obtained on the condition of many other glaciers. These data indicate that during 1957 glaciers were generally thickening and advancing in Washington and perhaps in Oregon; were thinning slightly in Montana, and were retreating in California. The summer of 1958 was one of exceptional ablation and caused a marked volume reduction in all glaciers measured as well as a decrease in the numbers of glaciers advancing. The 1958-59 budget year was slightly favorable for the growth of glaciers but there is no indication that a cycle of advancing glaciers has resumed.

The existence of glaciers in the United States south of Alaska has been known for nearly a hundred years. However, an accurate count of the total amount of glacier ice has not been possible until recent years because of the large amount that occurs in relatively inaccessible and seldom-visited areas. Several attempts at a glacier census have been made (Russell, 1858; Wentworth and Delo, 1931). More recently, Field and others (1958) summarized existing knowledge on the subject.

With the advent of the International Geophysical Year, it seemed appropriate to (1) learn more exactly how much glacier ice presently exists and where it is located and (2) determine the present condition of these glaciers. A cooperative program to obtain this information was authorized by the Technical Panel on Glaciology of the U.S. National Committee for the International Geophysical Year. Several government agencies and universities cooperated by undertaking new projects or modifying existing projects so that the data could be interrelated. This article represents a preliminary summary of the results obtained during the International Geophysical Year 1957-58, and the year following, known as International Geophysical Cooperation 1958-59. The objective of this article is to summarize the pertinent data that were obtained so they can be related with other results collected during the IGY-IGC observation period. No attempt is made to present a detailed analysis of these data.

This report could not have been possible without the excellent cooperation of several organizations and persons, who are mentioned as their data are presented.

1. DISTRIBUTION OF GLACIERS

New maps plotted by the Forest Service and the Geological Survey, new aerial photography by the Geological Survey, the Forest Service, and the University of Washington, and especially a study of the Northern Cascades by A.S. Post of the University of Washington, and the Geological Survey permitted compilation by the author of data on glacier sizes, numbers, and distribution. Data gathered by Dysco (1952) and Phillips on Rocky Mountain and Oregon glaciers, respectively, added greatly to this compilation. The sizes of all larger glaciers and more than half of the

smaller glaciers were measured by planimeter on the new maps. In order to portray the size-distribution of glaciers and to permit the more rapid sizing of the remaining unmeasured glaciers, an arbitrary scale of glacier sizes was defined. The average area of glaciers within each class was determined from a sample of 264 measured glaciers. The class limits and the measured average areas are given in table 1.

TABLE 1
Glacier size class limits and average areas

Class	Glacier area limits km ²	Average area within each class km ²
I	Less than 0.5	0.169
II	0.5 - 1	.73
III	1 - 2	1.42
IV	2 - 4	2.97
V	4 - 8	5.11
VI	More than 8	9.48

For each glacierized area in the United States south of Alaska, data are presented on numbers, sizes, and total areas of glaciers in table 2. Geographic variation in the mean altitudes of glaciers lends insight into the variations in climatic environment. However, in any given region it was found that the mean altitude of a group of glaciers was a direct function of the average size of the glaciers in that group. Because all glacierized regions contain Class I glaciers, mean altitudes of these smallest glaciers only are given in table 2. The locations of the glacierized areas are presented in figure 1. The latitudinal variation of mean altitude along the Cascade Mountains-Sierra Nevada system and the Rocky Mountain system are shown in figure 2. Locations of glaciers and geographic variations in mean elevation of the glaciers in Washington State are given in figure 3.

The total number of glaciers listed in table 2 is nearly 1,000, and they cover an area of more than 500 km². About 77 percent of this ice-covered area occurs in the State of Washington. By estimating reasonable average thicknesses for glaciers in each of the size classes and summing, we estimate a total volume of ice of 65 km³ (53×10^6 acre-feet). Assuming that the average yearly ablation is 4 m of water, these glaciers contribute about $2,000 \times 10^6$ m³ (1.7×10^6 acre-feet) of water to streamflow in the West during the summer months.

Most (79 percent) of the glaciers are tiny (less than 0.5 km²) masses of ice nestled in protected cirques. These smallest glaciers aggregate 26 percent of the total area of ice, and but 10 percent of the estimated total volume of ice. Only in the Olympic Mountains, the Northern Cascade Mountains, on Mt. Rainier and Mt. Adams in Washington, and in the Wind River Range of Wyoming do glaciers larger than 2 km² in area occur. Most of these larger glaciers are of the valley type, but large cirque glaciers are not uncommon. The largest single glacier is Emmons, on Mt. Rainier, which is 6.9 km long and 10.7 km² in area. However, it is exceeded in size by the Carbon-Russell Glacier system (essentially a single trunk glacier fed by two tributaries) which is 9.7 km long and 13.0 km² in area, and is also on Mt. Rainier.

The geographic variation in mean altitudes shows a good qualitative relation to precipitation and latitude. Glaciers occur at the lowest altitudes in northwestern Washington State, where huge yearly precipitation totals (more than 5 m) are occa-

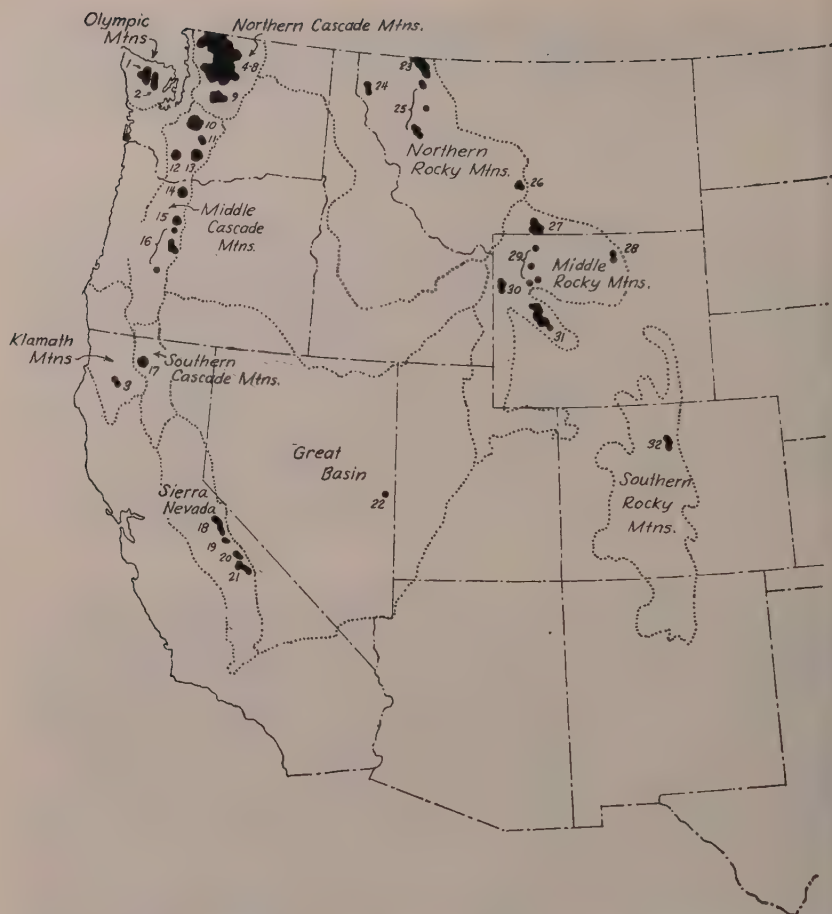
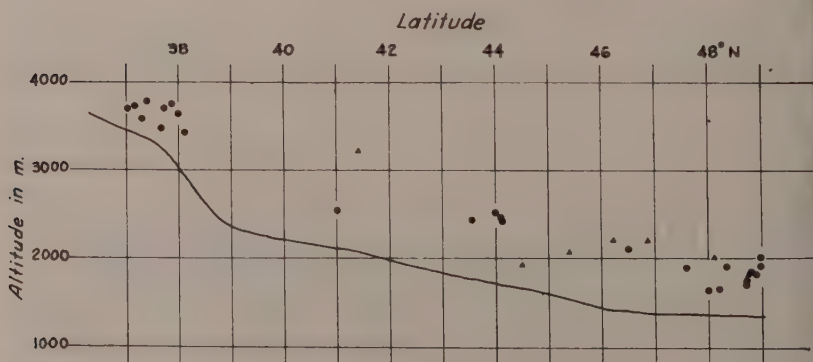


Fig. 1 — Map of Western United States, showing locations of glacierized areas. Dotted lines enclose those physiographic provinces which contain glaciers. Numbers are keyed to the glacier areas listed in table 2.



(a)

Fig. 2

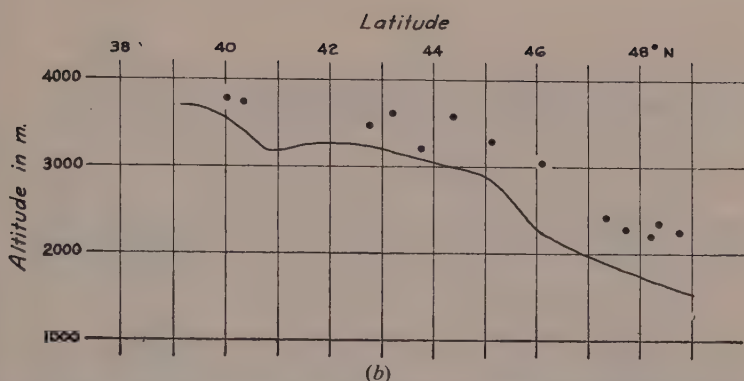


Fig. 2 — Graphs showing mean altitudes of Class I glaciers as a function of latitude in the Cascade Mountains and the Sierra Nevada (a) and Rocky Mountains (b). Each dot or triangle represents the average altitude of a group of from 2 to 63 glaciers. Glaciers on the seven major Cascade Volcanoes are indicated by triangles, other glacier areas are indicated by dots. The solid line indicates the altitude of Pleistocene cirque floors, taken from Flint (1957, p. 309).

sionally recorded. Glaciers occur at higher elevations further south and further inland. In general present-day small glaciers occur at altitudes from 300 to 600 m above the altitudes of Pleistocene cirque floors. If the mean annual precipitation had not changed, this would imply that the present-day summer climate is of the order of 2-4°C warmer than during the Ice Ages. Present-day glaciers occur only slightly above the Pleistocene cirque floors in the most southerly latitudes.

The vertical gradient in net accumulation (Shumskii, 1947) has been measured on only three glaciers in the conterminous United States. Values range from 15×10^{-3} (South Cascade Glacier, Washington) to 11×10^{-3} (Dinwoody Glacier, Wind River Range, Wyoming). These values indicate a relatively maritime climate. With the possible exception of glaciers in the Sierra Nevada, there is reason to believe that no highly continental-type glaciers occur in the conterminous United States.

2. VARIATIONS IN THE GLACIERS, 1956-59

Detailed regimen studies were made during the International Geophysical Year on Blue Glacier (La Chapelle, 1959) and South Cascade Glacier, both in Washington. Repeated topographic surveys provide data on the growth or shrinkage of four additional glaciers during the period 1956-59. Additional incomplete quantitative or qualitative data have been obtained on a large number of glaciers. Pertinent quantitative data are summarized in table 3.

An attempt has been made in table 3 to compare the relative intensity of ablation processes from glacier to glacier, by listing the measured or computed surface ablation rate in late summer. This was not measured on Nisqually, Grinnell, or Sperry Glacier, but the rate of lowering of the ice surface relative to sea level was measured. Ablation (V_a), lowering or raising of the ice surface (V_s) and the flow of ice normal to the surface (V_d) are related as follows (Meier, 1960) :

$$V_a = V_s - V_d$$

In this equation all components can be resolved either perpendicular to the surface or in a vertical direction, and velocities directed upward are considered positive. We assume that these glaciers were nearly in equilibrium ($V_s = 0$ over one budget year) and that V_d for one month was equal to $1/12 V_d$ for one year. The total yearly

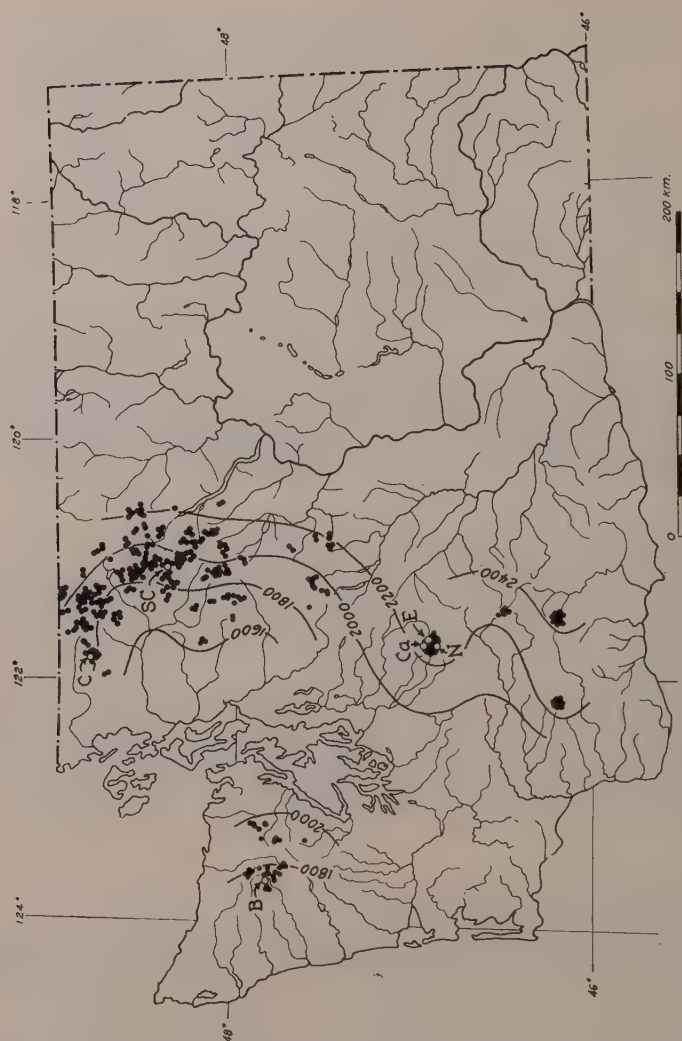


Fig. 3 — Location of glaciers in the State of Washington. Contour lines indicate mean altitude of Class I glaciers. Glaciers identified by letter are as follows:

- B Blue Glacier
- C Coleman Glacier
- Ca Carbon Glacier
- E Emmons Glacier
- N Nisqually Glacier
- SC South Cascade Glacier

ablation for each glacier was estimated, and $1/12$ of this value was assumed to represent $1/12$ of V_d and was added to the measured V_s during the month August 15 to September 15 in order to obtain V_a for this period. The corrections due to V_d were not large (see Table 3, notes 5 and 6).

TABLE 2

Distribution of glaciers in the conterminous United States, 1958

Location	Latitude	Size Class						Total number	Total area km ²	Altitude Class I glaciers m
		I	II	III	IV	V	VI			
1. Western Olympic Mountains	47°50'	22	8	3	1			36	27.0	1710
2. Eastern Olympic Mountains	47°50'	23	1	1		2		25	6.0	1880
3. Salmon-Trinity Mountains	41°00'	2						2	.3	2600*
4. Mt. Baker area	48°47'	17	5	1	1	5	1	30	55.5	1800*
5. Shuksan-Bacon-Challenger-Redoubt area	48°50'	126	11	4	6			147	52.6	1850
6. Dome-Eldorado area	48°25'	94	21	11	2	6		134	82.0	1930
7. Northeastern part, Northern Cascade Mountains	48°40'	39	2	1				42	7.5	1980*
8. Glacier Peak-Bonanza area	48°06'	74	10	7	5			96	39.9	2100
9. Southern part, Northern Cascade Mountains	47°35'	62	7	1				70	14.2	1950*
10. Mt. Rainier	46°52'	18	3	5	7	5	3	41	87.8	2230
11. Goat Rocks area	46°30'	9						9	1.5	2100*
12. Mt. St. Helens	46°12'	17	4					21	7.3	2220
13. Mt. Adams	46°13'	14	4	4				23*	16.1*	2500*
14. Mt. Hood	45°22'	4	3	5		1		12	9.9	2100*
15. Mt. Jefferson	44°41'	2	2	1				5	3.2	?
16. Three Sisters area	44°08'	15	5	1				21	7.6	2440
17. Mt. Shasta	41°25'	3	3	2				8	5.5	3200*
18. Yosemite National Park	38°00'	21						21	3.5	3590

TABLE 2 (Continued)

Distribution of glaciers in the conterminous United States, 1958

Location	Latitude	Size Class						Total number	Total area km ²	Altitude Class I glaciers m
		I	II	III	IV	V	VI			
19. Ritter-Minarets area	37°40'	4						4	0.7	3530
20. Abbot-Humphreys area	37°20'	8						8	1.4	3730
21. Goethe-Goddard-Palisade area	37°06'	36		1				37	7.5	3730
22. Wheeler Peak	38°59'	1						1	.2	3600?
23. Glacier National Park	48°45'	47	4	2				53	13.8	2280
24. Cabinet Range	48°13'	3						3	.5	2200
25. Flathead-Mission-Swan Ranges	47°40'	7						7*	1.2*	2430*
26. Crazy Mountains	46°05'	3*						3*	.5*	3050*
27. Beartooth Mountains	45°07'	34	5	1				40	10.8	3300
28. Big Horn Mountains	44°23'	2						2	.3	3600*
29. Absaroka Range	44°00'	4*						4*	.7*	?
30. Teton Mountains	43°45'	12						12	2.0	3200
31. Wind River Range	43°10'	45	6	5	5	2		63	44.5	3620
32. Rocky Mountain Park-Front-Range	40°03'	10						10	1.7	3800
Total number		778	104	56	27	21	4	990		
Area, km ²		132	77	79	80	107	38	513		
Assumed thickness, m		50	75	100	150	200	300			
Volume, km ³		6.6	5.7	8.0	12.0	21.4	11.4	65		

(*). These values represent estimates.

Glacier	Surface rise in m			Advance of terminus in m		Gross accumulation in m of water		Ablation rate Aug. 15-Sept. 15 in cm/day of water		Measured precipitation in mm		Note
	1956-57	1957-58	1958-59	1956-57	1957-58	1957-58	1958-59	1957-58	1958-59	1957-58	1958-59	
Blue	+ 0.6	- 2.1	+ 0.1	+ 58	+ 49	3.5	2.8	3.4	2.0	3780	3560	1
Coleman		- 7.0	+ 0.3									2
South Cascade		- 2.7	+ 0.8									3
Carbon												4
Nisqually	+ 6.1	- 7.9	+ 5.5	- 4.0	- 5.5	2.4	3.7	4.4	3.7		5330	5
Griinnell	- 1.1	- 2.9	+ 1.3	+ 39	+ 42			10.8	3.5	2612	3603	6
Sperry	- 1.2	- 2.9	+ 1.5	Little change	+ 17.3			5.1	2.5	2470	3760	7

Notes:

1. Data on gross and net accumulation and ablation rate obtained by E. R. La Chapelle (1959), University of Washington (IGY Project 4.3). Data on terminus advance supplied by the National Park Service (G. D. Gallison, personal communication April 1, 1960). Olympic National Park. Data on surface rise computed from measured net accumulation data assuming an area-averaged density at 2100 m altitude.
2. Data supplied by A. E. Harrison (personal communications February 1, 1959, and April 1, 1960), University of Washington. Data on surface rise apply only to the area below an altitude of 2400 m, about 58 percent of the total area of the Coleman-Roosevelt Glacier system. an area-averaged density at the surface of 0.8 gms/cm³.
3. Data obtained by the U.S. Geological Survey (M. F. Meier). Data on surface rise computed from net accumulation data assuming an area-averaged density at the surface of 0.8 gms/cm³ at the end of the ablation season. Glacier terminates in a deep lake, and the position data refer to the area below an altitude of 2100 m, about 17 percent of the total area of the Nisqually-Wilson Glacier system.
4. Data obtained by the National Park Service, Mt. Rainier National Park (Bender, 1958, 1959). Data on surface rise computed assuming $V_d = +1.7$ cm/day, from surface rise data obtained on a cross profile at 1840 m altitude. Precipitation measured at 1692 m altitude. along three radial profiles extending from the terminus almost to the head of the glacier.
5. Data obtained by U.S. Geological Survey (Johnson, 1958, 1960) in cooperation with National Park Service. Surface rise data apply area and $V_d = 0$ for 2/3 area. Precipitation measured at an altitude of 1881 m.
6. Data obtained by U.S. Geological Survey (Johnson, 1958, 1960) in cooperation with the National Park Service, along one transverse and two longitudinal profiles from the terminus almost to the head of the glacier.

In the Northwest the 1957-58 budget year was characterized by an early spring and a warm summer that was unusually long. The high ablation rates in August and September are attributed partly to a high incoming energy flux and partly to the abnormally low albedo of the glaciers due to the length of the ablation season.

The 1958-59 budget year was characterized by a relatively heavy winter accumulation of snow, a cool and wet spring, and frequent storms during the summer.

The data in table 3 reveal that during 1956-57 the Coleman and Nisqually tongues grew while the Grinnell and Sperry Glaciers declined slightly. All of the measured glaciers showed an appreciable reduction in volume during the year 1957-58. During 1958-59 all of the glaciers showed slight growth. Note that the data on advance of the termini do not correlate well with the overall volume change data. This is principally because the dynamic adjustment of valley glaciers (e.g. Carbon and Nisqually) to climatic change appears to take place by the development and propagation of kinematic waves (Weertman, 1958). These waves may arrive at the terminus several years after their initiation. There are suggestions in the 1959 survey data that two of these waves (Giles, G.C., personal communication April 18, 1960) may have been in progress on Nisqually Glacier.

It is perhaps surprising that the ablation rate data from the different glaciers are not markedly dissimilar, after allowing for some differences in the locations of sampling points. The similarity of local climatic environment is also suggested by the precipitation data. Thus, Grinnell Glacier appears to be in almost the same type of local climatic environment as Blue Glacier. Blue Glacier is but 52 km from the Pacific Ocean (a principal source of moisture-laden air), and occurs near verdant rain forests. Grinnell Glacier is 800 km from the ocean, and is separated from it by vast expanses of semiarid land.

Qualitative data on the variations of glaciers in the Northern Cascade Mountains, Washington, have been summarized by LaChapelle (1960). He reported on about 26 glaciers. During the 1956-57 budget year, 18 of these were actively advancing and 4 were retreating. During 1957-58, 9 were advancing and 5 were retreating. During 1958-59, only 4 were advancing whereas 12 were retreating.

Only three glaciers on Mt. Rainier, Washington, were observed during this period. Data on two of these are presented in table 3. Emmons Glacier, on the same mountain, advanced continuously from 1956 to 1959. On Mt. St. Helens, one glacier is known to have advanced at least until 1958, another was apparently retreating during this period, and no others have been studied. Few, if any, of the glaciers on Mt. Adams show evidence of reactivation, thickening, or advance; many show evidence of thinning and recession.

The lowest extremity of Eliot Glacier, on Mt. Hood, Oregon, has thinned continuously since before 1956; a profile at a higher elevation showed no appreciable change from 1955 to 1957 but the ice became thicker in 1958 and 1959; and points further up the glacier showed increases since 1957 ranging from 1.5 m to 6 m per year according to K.N. Phillips (personal communication to J.B. Case, September 29, 1959). Several other glaciers, as far south as Three Sisters, showed evidence of renewed activity in 1957.

Lyell Glacier, in Yosemite National Park, California, thinned 0.57 m from 1956 to 1957, according to surveys along a cross profile by personnel of the National Park Service and the Geological Survey (Garrison, et al 1957). Further south in the Sierra Nevada, O. Kehrlein (personal communication to W.O. Field, December 10, 1959), reported that Palisade and Powell Glaciers also were retreating in 1958, but Howell Glacier was holding its own.

Arapaho Glacier, Colorado, was probably retreating from 1957 to 1959, according to H.A. Waldrop (personal communication February 2, 1960).

It appears that the recent advance of glaciers discussed by Hubley (1956) was largely confined to the States of Washington and Oregon, possibly extending to Mt. Shasta in California (O. Kehrlein, personal communication to W.O. Field, December 10, 1959). Glaciers in the Sierra Nevada have apparently continued to waste away, whereas those in Glacier National Park, Montana, have remained very close to an equilibrium condition. The most spectacular advances occurred in the Cascade Mountains from Mt. Rainier north to the Canadian border, especially on Mt. Baker and on the Eldorado massif.

This advance was slowed appreciably by an abnormally heavy ablation season in 1958. Although conditions slightly favored glacier growth in 1958-59, all indications suggest that the recent cycle of advancing glaciers has not yet resumed and may be near an end.

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PARTICULARITES MORPHOLOGIQUES ET TENDENCE D'EXTENSION DE LA GLACIATION DU PAMIR NORD-OUEST

V. SOUSLOV

SUMMARY

The glaciological research carried out by us within the framework of the IGY and the IGS have enabled us to determine the specific peculiarities in ice formation in the high-altitude zones of the North-West Pamirs and also to establish a number of regularities in ice formation on the example of the Fedchenko glacier basin.

The morphology of the region of ice-formation is one of the major factors determining the direction and course of glaciological processes.

The peculiarity of the morphological structure of the glacier region is seen in the absence of the characteristic expansion of valleys in the neve region, in the sharp asymmetry of the glacier tributaries and the peculiar distribution of the velocities in the movement of the glaciers.

An important feature in the Fedchenko glacier system is the simultaneous development of two diametrically opposite processes — that of the advance and retreat of the glaciers.

The reduction in the area of glaciers is the governing direction in the development of present-day ice formation in the high altitude zone in the North-West Pamirs. Against this background, the advance of individual glaciers is of secondary significance.

RÉSUMÉ

Les recherches glaciologiques entreprises dans le cadre de l'A. G. I. nous permirent de découvrir les particularités spécifiques de la glaciation des régions de haute montagne du Pamir Nord-Ouest et de formuler certaines lois de son extension sur l'exemple du bassin du glacier de Fedtchenko.

Un des facteurs essentiels dictant l'orientation et la marche des processus glaciologiques est la morphologie de la région de glaciation.

La structure morphologique particulière de la région des glaciers se manifeste par l'absence de l'extension caractéristique des vallées dans la région du névé, par l'asymétrie marquée de la disposition des affluents des glaciers, la particularité de la distribution des vitesses de progression des glaciers etc.

Une importante particularité du système du glacier Fedtchenko est le développement simultané de deux processus contraires — recul et avancement des glaciers.

La réduction de la surface des glaciers est l'orientation essentielle du développement de la glaciation actuelle des zones de haute du Pamir Nord-Ouest. Sur ce fond, l'avancement de certains glaciers n'a qu'une importance subordonnée.

La présente communication découle de l'analyse préalable des résultats des observations effectuées dans le Pamir par l'expédition glaciologique de l'Institut de Mathématiques de l'Académie des Sciences de la RSS d'Ouzbékistan pendant la durée de l'Année Internationale Géophysique (1957-1959). Elle a pour but d'éclaircir certaines questions de morphologie et celle du développement de la glaciation dans le bassin du glacier Fedtchenko.

Les recherches glaciologiques ont été menées sur une superficie de plus de 1000 kilomètres carrés, aussi bien dans des stations fixes d'hivernage, dont l'une fut installée dans la région du névé du glacier Fedtchenko à une altitude d'environ 5000 mètres que par des détachements dont l'itinéraire avait été fixé à l'avance. On y a utilisé la prospection sismique, la photographie aérienne ainsi que des levées photographiques et au théodolite. Ces dernières furent effectuées par un groupe de géodésiens allemands (République Démocratique Allemande), sous la direction de l'ingénieur diplômé G. Diettrich, en collaboration avec des savants ouzbeks.

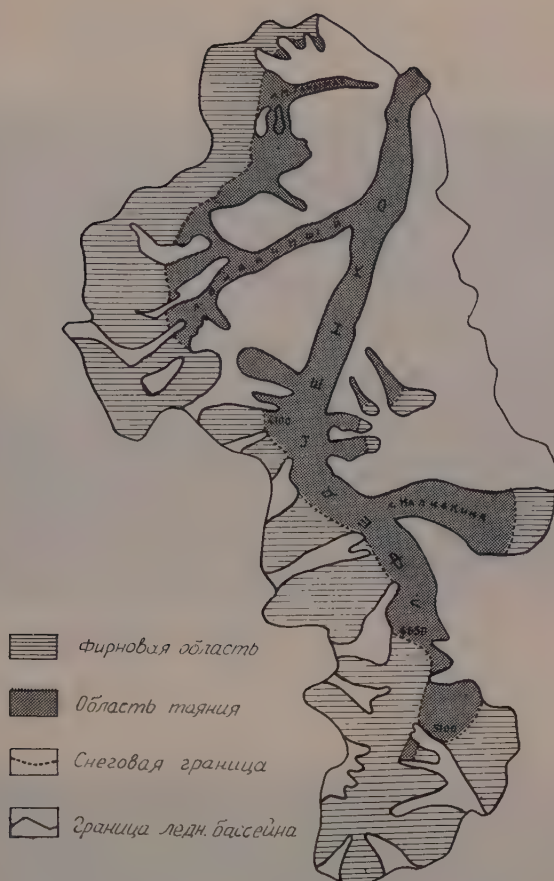


Fig. 1 — Schéma du bassin du glacier Fedtchenko

Il convient de noter que depuis l'époque de la deuxième Année Internationale polaire (1932-1933) aucune étude glaciologique tant soit peu importante n'avait été faite dans cette région, ce qui a provoqué certaines difficultés dans l'appréciation du caractère des processus de glaciation. Malgré cela, la confrontation des données des observations obtenues au cours de l'Année Internationale Géophysique dans le bassin du glacier Fedtchenko avec les résultats des recherches des années précédentes s'est avérée particulièrement intéressante et a permis de révéler certaines tendances dans le développement de la glaciation actuelle du Pamir du nord-ouest.

La situation d'élévation dominante des chaînes montagneuses de l'Académie des Sciences, de Pierre Ier, de Darvaz, de Vantch, de Yazgoulem et de Tanymass du nord, dont les sommets atteignent 7000 m et plus, au-dessus du niveau de la mer, en combinaison avec les particularités climatiques de cette région, a créé des conditions favorables à l'existence d'une glaciation énorme. Toutefois, ces considérations générales doivent être encore développées et étayées par de nouvelles preuves en rapport avec les conditions concrètes du bassin du glacier Fedtchenko qui se distingue des autres glaciers de la zone tempérée de la Terre par ses dimensions. Ceci s'avère



Fig. 2 — Partie terminale du glacier Fedtchenko



Fig. 3 — Partie médiane du glacier Fedtchenko



Fig. 4 — Loge de névé Kachal-Aiak



Fig. 5 — L'amont du glacier Fedtchenko



Fig. 6 — Le glacier Ouloubek. Zone de contact entre la glace enfouie et la glace active.

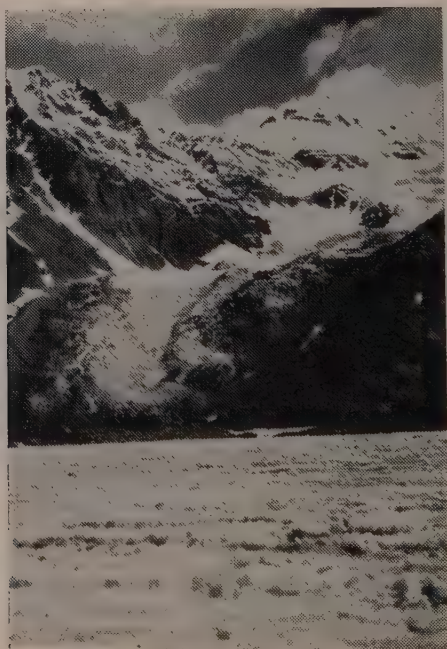


Fig. 7 — Le glacier Bezymianny

d'autant plus important que des causes purement climatiques, parfois, non seulement n'expliquent pas, mais souvent ne sont pas en harmonie avec la glaciation existante.

Le glacier Fedtchenko, qui suit pratiquement le méridien du sud au nord sur une distance de 77 kilomètres, remplit le fond d'une grande vallée et se termine à une altitude de 2900 m. Sa largeur varie de deux à quatre kilomètres et son épaisseur maximum atteint 1000 mètres environ. Nous avons considéré comme début du glacier la Grande auge synclinale de névé qui se trouve dans le massif de Yazgoulem, à une altitude de près de 6000 m. Quarante-deux grands glaciers-affluents dont la surface totale atteint 992 km² font partie de l'ensemble du glacier Fedtchenko.

Comme on le sait, les plissements montagneux du Pamir représentent tout un système de puissantes chaînes latitudinales parallèles, ce qui a marqué d'un sceau particulier la morphologie et la distribution de la glaciation actuelle, qui coïncide surtout avec celle des vallées montagneuses ayant la même orientation. Sur ce fond, la vallée du glacier Fedtchenko, qui s'étend sur près de cent kilomètres à travers les massifs montagneux, présente une anomalie dont la nature n'était pas tout à fait compréhensible jusqu'à ces temps derniers.

Les nouvelles données obtenues au cours des recherches éclairent d'un jour nouveau les origines et l'emplacement particulier de la vallée du glacier Fedtchenko dans le schéma tectonique général du Pamir.

Les recherches détaillées géologo-géomorphologiques combinées avec la prospection sismique confirment les notions sur l'origine érosive de la vallée du glacier Fedtchenko. Il fut établi que sur toute la longueur de la vallée (à partir du village d'Altynmazar jusqu'à Yazgoulem), il n'existe pas de différence tant soit peu importante dans la structure géologique et la constitution des roches qui forment les bords de la vallée. L'influence des processus tectoniques sur la formation de la vallée s'est avérée insignifiante et est apparue dans l'estuaire du glacier Bivatchny seulement. C'est là qu'une fissure locale de la faille s'étendant du sud-ouest au nord-est le long du bord droit du glacier Bivatchny jusqu'au versant droit de la vallée du glacier Fedtchenko a été révélée. La partie nord du massif est effondrée et sa limite coïncide avec la direction de la vallée de l'affluent.

Une couche intermédiaire, découverte par les sismologues V.A. Pak et V.N. Yacovlev, qui sépare la glace des roches de base du lit, témoigne également en faveur de l'origine érosive de la vallée du glacier Fedtchenko. Cette couche peut être suivie distinctement tout au long de la vallée du glacier Fedtchenko et en partie dans la vallée du glacier Bivatchny d'après le changement brusque de la vitesse de propagation des ondes réfléchies. De l'avis de V.A. Pak et de V.N. Yacovlev, cette couche constitue un ensemble alluvial ou de moraine cimenté par la glace. Il est caractéristique que, dans la partie inférieure du glacier, cette couche atteint son épaisseur maximum, soit près de 700 m; dans la partie médiane — 370 m et en amont — près de 100 m. La propagation générale et la diminution uniforme de l'épaisseur du dépôt, au fur et à mesure de la montée vers l'amont de la vallée, témoigne davantage du développement important des processus d'accumulation fluviale à l'époque pré-glaciaire et du remplissage de la vallée par des alluvions, que de la nature glaciaire de ceux-ci. Dans ce dernier cas, il aurait fallu admettre de nombreuses époques glaciaires entrecoupées de périodes prolongées de réchauffement, ce qui n'est point confirmé par les données paléoglaciologiques (K.K. Markov).

Un facteur non moins important qui confirme notre point de vue sur l'origine de la vallée est l'uniformité exceptionnelle et la faible déclivité du profil longitudinal du lit du glacier, très proche du profil d'équilibre d'une vallée fluviale formée. Ce n'est que dans la région du confluent du glacier Bivatchny avec le glacier Fedtchenko et dans la partie terminale de ce dernier que se distingue une déclivité peu importante du lit vers l'amont de la vallée, ce qui s'explique par des raisons tectoniques.

Ainsi la liaison génétique de la vallée du glacier Fedtchenko avec le processus de l'érosion pré-glaciaire justifie son état d'anomalie dans le système des chaînes montagneuses du Pamir, et révèle dans une mesure importante l'échelle de la glaciation actuelle de l'amont de la rivière Moukhsou, déterminée par la large cuvette de son lit et par son exposition au nord.

Il n'est pas difficile de se convaincre de ce que l'orographie et l'exposition des versants ont une influence considérable sur l'emplacement de la limite des neiges et sur le rapport des régions morphologiques principales du glacier. Le niveau de la limite des neiges, qui fut déterminé par diverses méthodes pour le glacier Fedtchenko, approche de 450 mètres. Toutefois, les observations de ces dernières années ont démontré que ce niveau ne correspond à la position réelle de la limite des neiges que sur l'alignement principal du glacier Fedtchenko, dans la région du glacier Tanymassky Loskout et s'écarte considérablement de la marque médiane sur les glaciers-affluents importants, en rapport avec l'exposition de ces derniers. En même temps, l'amplitude de la déviation de la limite des neiges de sa position médiane dans l'alignement principal du glacier atteint 1000 m, ce qui n'est que de 340 m inférieur à la différence d'altitude maximum de la limite des neiges notée par R.D. Zabirov pour le Pamir tout entier.

La limite des neiges arrive à son niveau le plus élevé au glacier Vitkovsky et Dorofeyev, atteignant ici l'altitude de 5100 m. Son niveau inférieur est enregistré dans l'estuaire du glacier Kachal-Aïak et il passe le long d'un axe horizontal sur 4100 m. Ainsi la position moyenne de la limite des neiges de tout le bassin du glacier Fedtchenko reste voisine du niveau de 4600 m. Toutefois, ses écarts importants de ce niveau influencent considérablement la répartition des régions morphologiques principales du glacier.

Ainsi qu'il apparaît du schéma, la région du névé du glacier Fedtchenko est nettement asymétrique par rapport à son tronc principal et comprend l'amont de la vallée et la majeure partie de la pente est de la chaîne de l'Académie des Sciences, jusqu'à la latitude du col Kachal-Aïak. Les fleuves de névé qui coulent de cet endroit le long des gorges importantes et qui sont dénommées glaciers sans raison valable, simplement à cause de leur ressemblance apparente avec ces derniers, se trouvent tout entier dans la région de l'accumulation. Ceci dit, il serait plus juste de dénommer les glaciers Kachal-Aïak, Elena Rozmirovitch, Académie des Sciences — auges de névé, ce qui correspondrait à leurs caractères morphologiques. Toutefois, en ce qui concerne les glaciers Nalivkine et Dorofeyev qui sont exposés à l'ouest, il convient de garder le terme de glacier, car dans leur cas nous avons une délimitation très nette des régions d'accumulation et d'écoulement.

L'asymétrie de la région du névé et l'amplitude importante de fluctuation des niveaux de la limite des neiges dans le cadre d'un système de glaciers sont conditionnées par l'orographie et témoignent de la présence de grands contrastes dans les conditions climatiques de la région. La dépression importante de la ligne des neiges dans la région de l'auge du névé Kachal-Aïak (4100 m) peut être considérée comme étant dépendante de l'influence refroidissante des grands massifs montagneux de Darvaz et des conditions favorables d'accumulation de précipitations du versant nord-est sous-le-vent et du col Kachal-Aïak. Dans «l'ombre de vent» par rapport à la direction sud-ouest des courants aériens qui prévalent ici, se trouvent les auges du névé E. Rozmirovitch et de l'Académie des Sciences, dans lesquelles l'accumulation des précipitations a principalement lieu pour le compte du déplacement de la neige.

Les données obtenues au cours de l'Année Internationale Géophysique ont permis de tirer une certaine appréciation quantitative de la condition dynamique du glacier Fedtchenko qui est caractérisée d'après la méthode de P.A. Choumsky par la somme de gradients verticaux d'accumulation et d'ablation. Il a été établi qu'en 1959 la ligne des neiges passait le long du tronc principal du glacier au niveau de 4750 m et que la

La valeur d'intensité d'accumulation-ablation était à cet endroit de 1484 mm. En montant de 76 m au-dessus de la limite des neiges le gradient vertical d'accumulation est reconnu être de 5,3 mm par mètre, et le gradient vertical d'ablation de 1490 mm par mètre. Ainsi l'augmentation de l'accroissement annuel des précipitations sur les glaces, autrement dit l'énergie de la glaciation, exprimée par la somme des gradients, est égale à 8,3 mm par mètre.

La comparaison de cette valeur avec les données sur d'autres régions glaciaires et en particulier avec celles des glaciers de la Terre du Nord-Est — de 0,5 mm par mètre; du glacier Vatnaökoul (Islande) — de 6,5 mm par m; du glacier du Rhône (Alpes) — de 12,4 mm par mètre, fait ressortir le degré suffisamment grand d'activité du glacier Fedtchenko. Toutefois, en présence des expressions quantitatives identiques de l'énergie de glaciation, les glaciers peuvent se trouver aussi bien dans une période de progression que dans une période de régression. La valeur de l'énergie de glaciation dans ce cas particulier démontre simplement une stabilité plus ou moins grande de la position de la limite des neiges et on peut en tenir compte dans l'appréciation de l'état de la glaciation.

Quelle est donc la tendance du développement de la glaciation actuelle de la vallée de la rivière Moukhsou? L'analyse des données sur l'état de la partie terminale du glacier Fedtchenko au cours de la période s'étalant de 1913 à 1957, effectuée par P. Tchertanov, démontre qu'au cours des derniers 45 ans la «langue» du glacier a cessé de se contracter et de reculer. Pendant ce temps, la surface de la glace fondue a égalé 2,5 km². Les grands glaciers affluents M. Tanymass, Kocinkenko, Alert et autres, qui ont déjà perdu leur liaison directe avec le glacier Fedtchenko et sont entrés dans leurs vallées sur une distance de 1 à 2,5 kilomètres se trouvent également dans une phase régressive. Le grand glacier Bivatchny a beaucoup diminué de taille, et sa partie estuaire tend également à se séparer du tronçon principal du glacier Fedtchenko. Sur le fond de la diminution générale de la glaciation du bassin, certains glaciers tels que ceux d'Ouloubek, de Kalinine et Bezymianny se trouvent actuellement dans un stade d'activation.

Le glacier Ouloubek, situé dans le système de la chaîne Kyz-Kourgan, est éloigné de 2,5 km du glacier Fedtchenko et à l'heure actuelle il progresse énergiquement, la langue du glacier qui est située à une altitude de 4130 m et affecte une forme convexe caractéristique en témoigne d'une façon convaincante. La surface de la partie linguale du glacier manque presque totalement de moraine et est couverte d'un réseau serré de failles qui doivent leur origine aux mouvements énergiques de la glace.

La preuve la plus convaincante de la recrudescence de l'activité du glacier Ouloubek est le fait de la montée de la jeune glace du névé sur l'épaisseur glaciaire au-dessus de moraines qui est d'une formation plus ancienne. La dénudation naturelle sur le long du bord gauche du glacier permet de suivre ce contact sur une distance de 2 km. La couche enterrée de la glace foncée émerge en coin sur une distance de 100 m de la partie frontale du glacier. Tout ceci témoigne de ce que le glacier Ouloubek, qui dans un passé récent se trouvait dans une phase de recul, progresse actuellement et a avancé sur une distance de 300 m au moins. Si on tient compte du fait que l'épaisseur sous-jacente de la glace souillée n'est pas immobile non plus, le mouvement du glacier apparaîtra comme étant encore plus considérable. Le glacier Kalinine (affluent gauche du glacier Bivatchny) peut servir d'autre exemple de la recrudescence de l'activité glaciaire dans cette région. Dans sa partie estuaire, on observe la montée d'un fleuve de glace du névé sur la «langue» du glacier couverte de moraines. Le fleuve glaciaire qui forme la seconde assise du glacier a avancé sur plus de 300 m au-delà de la région du névé.

La progression du glacier Bezymianny qui s'est formé sur le versant sud de la chaîne Kyz-Kourgan est particulièrement évidente. Les observations de ce glacier, qui ont été faites de 1956 à 1959, confirment que l'augmentation sensible de la langue

a eu lieu en 1956. A l'heure actuelle, le glacier continue à avancer et apparemment il se réunira bientôt avec le glacier Fedtchenko. Le mouvement des masses glaciaires a été suivi ici de leur empiètement partiel sur les restes de l'ancien glacier, ce qui a eu pour résultat que deux glaciers indépendants se sont formés dans la vallée, dont l'un a avancé de 300 à 350 m par rapport à l'autre.

Tous les cas de progression des glaciers préalablement cités se rapportent au bassin du glacier Fedtchenko et possèdent un caractère local. Ceci témoigne de la grande variété des conditions locales dont les manifestations dépendent, dans une certaine mesure, de certains changements d'ordre général. Ainsi par exemple, au cours des dix années allant de 1934 à 1945, la quantité moyenne des précipitations pendant les trois mois d'hiver a été de 254 mm dans la partie médiane du glacier Fedtchenko, tandis que dans la période de l'hiver de 1957-1958, la quantité moyenne des précipitations a atteint ici 407 mm. Ce n'est pas par hasard que c'est au cours de cette période que se rapporte l'avance maximum des glaciers que nous avons noté et dont les bassins doivent accumuler la plus grande quantité de précipitations atmosphériques solides, grâce à leur exposition au vent. D'un autre côté, l'absence de synchronisme dans les fluctuations de la dimension des glaciers du système Fedtchenko peut être justifiée par la différence importante de leur échelle. En effet, en règle générale les glaciers en progression ne sont pas grands et pour cette raison ils réagissent plus vite aux changements des conditions de leur alimentation.

En faisant un bilan partiel on peut noter ce qui suit :

1) L'un des facteurs primordiaux déterminant la direction et l'échelle des processus glaciologiques dans le bassin du glacier Fedtchenko consiste dans les particularités morphologiques de la région.

La situation d'anomalie de la vallée du glacier Fedtchenko, situé en travers de la direction des chaînes montagneuses du Pamir du nord-ouest, ne peut être justifiée par des raisons tectoniques. La liaison génétique de la vallée avec les processus de l'activité érosive préglaciaire apparaît comme la plus véridique, ce qui trouve une certaine répercussion dans le caractère de la glaciation actuelle du bassin.

2) Les particularités morphologiques de la région de glaciation exercent une grande influence sur la position de la limite des neiges. Dans le système du glacier Fedtchenko l'amplitude des oscillations de la limite des neiges est de 1000 m. En cela le rôle principal est joué par l'exposition des versants par rapport au soleil et la direction prépondérante des courants aériens.

3) La région du névé du glacier Fedtchenko est asymétrique et est déportée vers le nord-ouest de son tronc principal et c'est grâce à cela que certaines auges de névés se situent considérablement plus bas que le niveau moyen de la limite des neiges.

4) La haute activité du glacier, qui est caractérisé par son énergie de glaciation (8,3 mm par mètre), se trouve en contradiction avec son état régressif actuel.

5) Malgré la tendance générale à la contraction de la surface de glaciation au cours des derniers 45 ans, on observe dans le bassin du glacier Fedtchenko une reprise notable de l'activité de certains glaciers. Une telle absence de synchronisation dans les oscillations de la dimension des glaciers est en rapport avec les conditions morphologiques et climatiques particulièrement diverses des régions qui les alimentent.

6) Il apparaît comme très souhaitable que soient poursuivies les études systématiques des processus de glaciation dans le bassin du glacier Fedtchenko, aussi bien dans le sens de leur appréciation quantitative et de l'établissement de leurs pronostics que dans l'intérêt du développement de la théorie des phénomènes glaciaires.

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THE FEDTCHENKO GLACIER AND CLIMATE

V.A. GEORDIO, A.B. KAZANSKI, V.N. KOLESNIKOVA,
B.K. NOZDRUHINE, M.A. PETROSSIANTS

RÉSUMÉ

1. Les études sur le climat du bassin du glacier Fedtchenko avaient pour but d'éclaircir les particularités de l'accumulation de la neige et du névé sur le territoire de l'Asie Moyenne, malgré son climat aride.

2. Les observations furent faites par des stations météorologiques nouvellement installées dans la zone du névé (à l'altitude de 4900 m) et immédiatement avant l'extrémité de la langue du glacier, dans la vallée (à l'altitude de 2900 m).

Les observations étaient faites de jour et de nuit pendant toute la durée de l'Année Géophysique Internationale. De même, furent exécutées des observations épisodiques sur la fonte et le bilan thermique de la surface du glacier sur tout le corps du glacier pendant la fonte intense de la neige (de l'altitude de 5000 m à l'altitude de 3000 m).

3. On a obtenu une caractéristique climatique du bassin du glacier Fedtchenko. Des particularités du champ de la température et du vent ont été révélées. Ainsi, par exemple, l'amplitude diurne de la température dans la région du névé est de beaucoup plus grande que celles des stations dans la plaine et surtout que celle de l'atmosphère libre au même niveau.

Il n'existe pas de changements diurnes de la direction du vent dans la région du névé, c'est un vent d'écoulement qui prédomine. Près de la langue du glacier il existe une circulation « montagne-vallée » très nette. Cette circulation a un cours annuel.

Le vent d'écoulement de la région du névé est sujet à des pulsations de température (jusqu'à 6° en une demi-heure). Dans le cas d'existence au-dessus du Pamir d'une zone frontale planétaire à haute altitude les pulsations sont remplacées par un cours de température très régulier.

Les conditions de l'accumulation de la neige en amont du glacier sont expliquées. Les observations sur le bilan thermique de la surface du glacier, sur son déplacement, sur l'accumulation et la fonte de la matière ont permis de tirer des conclusions sur le rôle des eaux de fonte de la glace dans l'alimentation des rivières.

ABSTRACT

1. The study of climatic features of Glacier Fedtchenko basin was carried out in order to clear up the particularities of snow and ice accumulation upon the territory of Central Asia in the dry continental climate peculiar to it.

2. Observations were effectuated all round the whole day during the period of the I.G.Y. and M.G.S. at newly organized meteorological stations on the firn zone of the Fedchenko glacier (altitude 4900 m) and directly before the end of the glacier's tongue, in its valley (altitude 2900 m).

In the period of intensive snow-melting upon the whole glacier body (from an altitude of 5200 m to the altitude 3000 m) were carried out episodic observations over the thermal balance of the glacier surface and on melting.

3. Obtained was a climatic characteristic of the Fedchenko glacier's basin. Cleared up was a series of particularities in the field of temperature and of wind. For instance, the diurnal temperature amplitude in the firn region was found to be considerably greater than the amplitude on the flat country, and so much the more were amplitudes in free atmosphere at the same altitude.

In the firn region of the glacier, there is no diurnal change of wind directions; here dominates a flow-wind. At the glacier's tongue is clearly expressed the mountain-valley circulation, that has a yearly march. In the flow wind of the firn region there are temperature pulsations (up to 6° in half an hour). In presence over the Pamir of an evident P.V.F.Z., pulsations are substituted by a notably even temperature march, etc.

Explained are some conditions of snow accumulation in the upper reaches of the glacier. The observations carried out over the thermal balance on the glacier's surface, on its movement, on accumulation and melting of the substance, give possibility to elaborate a notion about the rôle of the glacier's feeding of rivers.

During the International Geophysical Year an expedition of the Institute of Mathematics of the Academy of Sciences of the Uzbek S.S.R. carried out glaciological research on the Fedtchenko Glacier—the biggest valley glacier on the earth.

The Fedtchenko Glacier is situated amidst mighty mountain heights in the north-west Pamirs. It takes its source at the junction of the Yazgulem and the Academy of Sciences mountain ridges. The glacier's direction is meridional in the main. Its southern end (upper) is located under the peaks of the Paris Commune (6,354 m), the 26 Commissars (6,834 m), the Revolution (6,974 m) and others. The northern end of the glacier (lower) stretches to the Seldar River Valley, 2,900 metres above sea level. The glacier is 77 km in length, and the width at its trunk comes to an average of 2.5 km. The area of ice and snow accumulation in the Fedtchenko Glacier basin comprises 992 km². The rivers flowing out of the basin feed the Amu-Darya—a big waterway Central Asia. The starting point for studying life on the glacier as a whole is the study of its climatic characteristics. In the main, stationary observations were conducted at three stations: at the observatory of Fedtchenko-1 Glacier (4,120 m) which was set up back in 1933 during the Second International Polar Year, and at the two stations—"Vitkovski Glacier" and "Fedchenko-2 Glacier", organized by the Academy of Sciences of the Uzbek S.S.R. during the International Geophysical Year. The "Vitovski Glacier" is situated in the central part of the neve region of the glacier. (4,900 m) and the "Fedtchenko-2 Glacier"—in the flood-lands of the Seldar River before the glacier's tongue (2,900 m).

1. RADIATION REGIME

Being situated between 39° 51' 40" and 38° 13' 27" Lat. (72° 23' 23" Long.) in an alpine zone, the Fedtchenko Glacier receives a large amount of warmth from the sun. For instance, in the upper reaches of the Fedtchenko Glacier, the total annual quantity of radiation comprises 194.3 $\frac{\text{kcal}}{\text{cm}^2}$. (The period under review being from October 1957 to October 1958). The mean annual latitudinal sum of this quantity for the region of the upper station amounts to approximately 133 $\frac{\text{kcal}}{\text{cm}^2}$. The greatest total annual quantity of radiation, known in literature, is to be observed in the tropics—up to 205 kcal/cm². (India: Poona, 18.5°). At the upper station, the greatest total quantity of radiation has been observed in June—26 kcal/cm² a month, and least warmth was received in December—6.8 kcal/cm². Included in the total radiation there is a great amount of diffused radiation. In winter and spring, at the upper station, it comprises 60-70% and in summer 20-30%.

The laws governing the changes in diffused radiation have been studied during different times of the year depending on the cloudiness on different parts of the glacier.

The mean annual value of the albedo is the greatest in the neve region—76%. The radiation balance in the neve zone on the average for the year is negative. In winter it is negative throughout the 24 hours of the day. The mean sums of the radiation balance for each month are as follows (1958-1959).

IX	X	XI	XII	I	II	III	IV	V	VI	VII
-0.2	-133.2	-109.8	-36.0	-80.4	-40.8	16.2	52.5	—	21.4	91.2

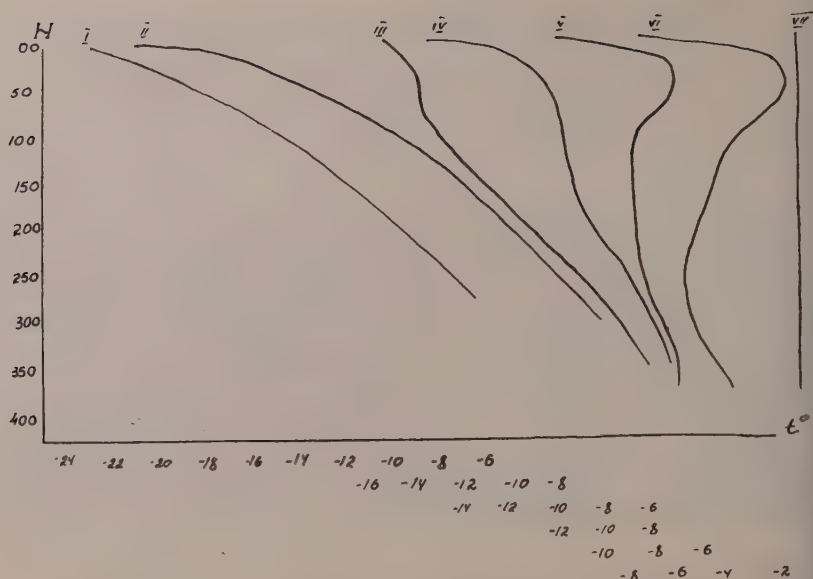


Fig. 1 — The mean monthly distribution of the temperature in the upper layers of the neve.

The great positive values of the radiation balance during separate warm periods of the year is explained by the pollution of the glacier's surface due to the penetration of dust from the southern and south-western regions, where dust storms occur at this time.

1.1. Temperature of the air

All the above-stated creates unusual conditions as compared with flat country for the formation of the temperature regime near the glacier's surface.

A notable peculiarity in the daily changes of the temperature in winter is the fact that the mean temperature curve for cloudy days is situated higher than the temperature curve for clear days. This applies to all three stations and is most strongly marked at the upper station. During the other seasons the effect is retained only partially at the upper station, which can be explained by the great degree of the cooling of the surface ice and snow due to radiation. (At the upper station loss of warmth due to radiation in the period under review came to 200 kcal/cm², a year. That is why cloudy weather, accompanied in winter by a new air mass, which has not undergone yet intensive cooling, causes a higher temperature of the air and the snow surface (it prevents the cooling of the surface). A comprehensive analysis of curves of maximum, mean daily and minimum temperatures has also shown that the amplitudes increase in periods when cold sets in and lessen in periods when it gets warmer when there is both extensive cloudiness and it rises correspondingly to the height of the station. An exception to this is the middle station, where the winds blowing from Kashal-Ayak ensure a good exchange of air, which results in the fact that the daily amplitudes of the temperature there are very small. In winter at the upper station

the slight recurrence of the values of the temperature comes to 15-19.9°C (there is a great recurrence also within the limits of 20-25°C), during the summer months—by 0.1 and 9°C, at the middle station —15.9-20° (there is a great recurrence also within the limits of —10.0 —15.0°C in winter, and at the lower —5 —10°C in winter, 15.0-15.0° in summer.

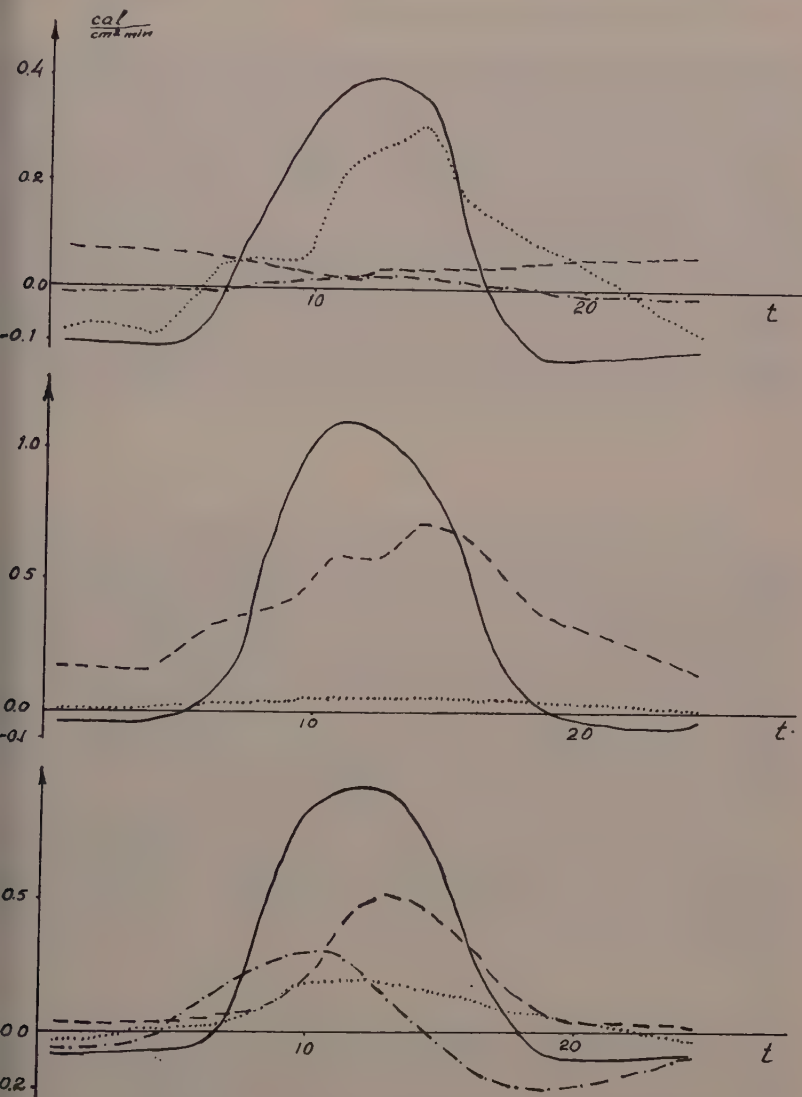


Fig. 2 — The mean daily course comprising the thermal balance on the Fedtchenko Glacier:

- a) the upper station,
- b) the middle station,
- c) the lower station.
- Radiating balance
- - - Turbulent stream
- - - Stream of heat, spent for evaporation
- . - . Stream of heat in the ground on account of heat conductivity.

By comparing with the regime of free atmosphere at corresponding levels, the cooling effect of the glacier, most strongly marked in summer, can be seen quite distinctly.

1.2. *Temperature of the surface layers of the neve*

The temperature of the glacier's surface in the space of 24 hours fluctuates sharply. On clear days in winter on an average of from -34° to -17°C ; on cloudy days—from -20° to 15°C ; in spring—from -23° to 1°C on clear days, from -15° to -17°C on cloudy days, in summer from -10° to 0°C on clear days and from -5° to 0°C on cloudy days. The daily fluctuation of the temperature quickly attenuates with depth. At a depth of 25 cm it comes to several tenths of degrees up to 1.2° . The greatest fluctuations occur in spring. The fluctuation of the temperature with depth is of quite a definite nature for each season. The greatest differences occur in the winter and summer distribution. (Fig. 2). In the winter time the average daily temperature is equal in the whole layer. With the coming of spring the rise of temperature spreads to ever deeper layers. In the summer time a sharp rise in temperature is observed due to the infiltration inside of the thawed snow. In July practically throughout the whole 24 hours of the day, beginning with a depth of 20, 30 cm, the temperatures are close to zero.

2. WIND

Circulation in the Fedtchenko Glacier basin is characterized by the following peculiarities. At the upper station throughout all the seasons a southern wind predominates, i.e., down over the glacier. This is a gusty wind, which predominates not only in the neve region, but over a considerable part of the glacier. The wind reaches the greatest velocity at night and at daybreak. A minimum velocity is to be observed in the daytime. At the middle station a south-western wind predominates the year round. Observations carried out on the glacier in the vicinity of the station show that there a southern wind predominates downwards over the glacier, i.e., the same gusty wind as at the upper station. At the lower station a very marked mountain-valley circulation is to be observed throughout the whole year. The mountain wind represents a diffusion of the gusty wind to the lower part of the glacier.

In the gusty wind of the neve region there is a pulsation of temperature (up several degrees in the course of half an hour). When there is present over the Pamir of a distinct planetary high-altitude frontal zone the pulsation is replaced by a remarkably even course of temperature. Certain other peculiarities of the wind regime in connection with large scale circulation have been established.

3. HUMIDITY

The humidity regime on the Fedtchenko Glacier has been studied in detail under different weather conditions depending on the time of the year and the day. Comparisons have been carried out with the regime of free atmosphere. At the upper station the greatest recurrence of values of relative humidity comes to 60-70% in the course of the year. There is a great recurrence of values of relative humidity also within the limits of 40-59%, especially in summer. At the middle station the relative humidity on the glacier is the highest during the period from October to May inclusive and its greatest recurrence comes to 80-100%. (The intrusion of cloudiness through

Kashal-Ayak). During the remainder months its greatest recurrence comes to 40-59%. At the lower station during the period from November to March inclusive its greatest recurrence comes to 40-59%, and the rest of the time to 20-39%.

3.1. Cloudiness

The region of the glacier is marked by considerable cloudiness. From November to May inclusive the number of cloudy days in the month comprises approximately 60-90%. At the upper and middle stations cloudiness intrusion frequently takes place at the height of the station. «The cloud cover settles to lower levels and then dissipates» During the summer the probability of a cloudy sky on the glacier is greatly lessened. During this time of the year, in most cases cumuli are to be observed there, formed as a result of local convections. At the lower station, where the surrounding mountain slopes and summits are mainly snowless and the atmosphere is sufficiently dry, considerably fewer clouds of these forms are to be observed. Cumuli, as a rule, never reach great size. Dense and heavy cumuli, as well as cumuli-pluvial clouds, appear in the region of the glacier during intrusion of cold fronts or soon after they pass.

On the Fedtchenko Glacier during certain periods rapid freezing is to be observed of the summits of the cumuli. This process reaches particular development in the neve region.

3.2. Precipitation

The accumulation of precipitation is connected with a whole series of orographic and climatic peculiarities of the glacier's basin. The Academy of Sciences ridge (mean height 5,757 m), stretching meridionally, bars from the glacier the humid western and south-western air masses, thanks to which part of the humidity is condensed on the western slope of the ridge. The slopes, separating the Fedtchenko Glacier valley from the west with the exception of its lower part, are almost completely covered with a mighty layer of ice and neve. On the other hand, the opposite slopes, except the extreme upper section, are bare. The western confluent of the Fedtchenko Glacier are in the main represented by neve basins, which can descend to 4,100 m. (Kashal-Ayak Glacier). But the eastern confluent, with the exception of the "Big Neve Basin" and the "Snezhnyi" Glacier are typical glaciers with independent regions of ablation and accumulation.

The neve border on the Dorofeyev Glacier—the most southern of them—rises to a height of 5,100 m. The altitude of the snow line on the body of the Fedtchenko Glacier fluctuates depending on the conditions of the year within the limits of 4,600-4,700 metres, and probably never rises higher than 4,750 metres, since from this level there begins a sharp increase in the thickness of the snow cover. Precipitation in the Fedtchenko Glacier valley is distributed extremely unevenly.

A maximum quantity of precipitation is to be noted in the region of the middle station. The lowest point of the Academy of Sciences ridge is the Kashal-Ayak pass (4,340 m). The presence of three other relatively low points in the upper reaches of the glacier—the Academy (4,805 m), the Abdukagor (5,058 m) and the Yazgulem (5,330 m) facilitates the accumulation of precipitation in the neve zone of the glacier. The wind plays an active part in the process of the accumulation of matter on the glacier. An enormous quantity of precipitation, deposited on the slopes surrounding the glacier, is blown off from them by the wind and is deposited directly on the main body of the glacier and its confluent. In the course of the year about 50% of the slope surfaces are deprived of their snow cover. The amount of snow drifted onto the

Monthly Deposit of Precipitation in mm.

St-ns	m-ths year	I	II	III	IV	V	VI	VII	VIII	IX	X	XI	XII	sum year
Fedtchenko-II	1958	10	14	12	50	16	8	22	2	0	4	12	42	192
	1959	4	13	27	6	34	0	0	0					
Fedtchenko-I	1958	173	76	214	380	188	65	56	20	28	138	166	284	1788
	1959	51	191	352	127	236	72	50	37					
Vitkovsky Glacier	1958	62	18	57	203	78	57	42	23	34	49	69	90	782
	1959	18	46	112	67	152	39	37	21					

body of the glacier in different periods sometimes comes to 30-40%. It should also be noted that the wind takes part in the redistribution of the snow over the glacier. For instance, an increase in the thickness of the snow cover is to be observed on the left bank. On the right bank the cover of snow is often blown off by the wind altogether. The wind, blowing at the lower part of the glacier, "presses itself" to the right side thanks to the western and south-western currents, penetrating through the pass in the Academy of Sciences ridge.

The seasonal snow on the Fedtchenko Glacier beneath the neve border melts and does not take part in feeding the glacier. The snow and neve cover in the zone of accumulation of precipitation (matter) is characterized by the following peculiarities: At the end of September and the middle of October a new snow cover is formed in the upper reaches of the glacier. Its density is 0.1-0.15 g/cm³. Under the action of the wind and under the load of its own weight the snow gradually compacts. By the beginning of December the density reaches (0.29-0.31 g/cm³). In the course of the winter the thickness of the snow increases more or less evenly. The increase of density is also slow. In spring the quantity of precipitation increases sharply and the thickness of the snow cover reaches its maximum value (by the end of May). At the Tanymas Lapen it reaches 180-200 cm, at the neve border—205-220 cm, in the region of the upper station—280-290 cm, at the Abdukagor pass—340-360 cm, and at the Big Neve Basin at an altitude of 5,156 m—430-450 cm. In June due to a sharp increase in the density of the snow cover its thickness decreases. The falling snow continues to increase its water content. At the upper station the maximum value of the water content in the annual layer of snow was checked at the end of June and found to be 1,150 mm. In the region of the neve border and the Tanymas Lapen at the end of May was 750-690 mm. At an altitude of 5,156 m, at the end of July—(1,650 mm). At the beginning of July, intensive snow melting begins in the neve region and continues until the middle of September. During this period the density of the snow increases from 0.42-0.44 to 0.5-0.6 g/cm³, and the daily decrease of matter calculated in water comprises: at the Tanymas Lapen—12.0 mm, at the neve border—11.4 mm, in the vicinity of the station—6.6 mm, at the Abdukagor pass—5.7 mm, at the Academy of Sciences of the Uzbek S.S.R. Glacier—1.5 mm and at an altitude of 5,156 m—0.5 mm. Not only the altitude of places above sea level influences the intensity of melting, but also the peculiarities of orography, microclimate, wind etc. Thus, in the neve zone on the main trunk of the glacier the melting at the summit decreases rapidly. From the neve border to the big neve basin (5,156 m) distance 13.5 km (area 33.3 km²) by July 1, 1959, the content of water in the snow on the average comprises 1,032 mm, which brings the general reserves of water in this area up to 34.4×10^6 m³. By September 7, the water content decreased to 805 mm, and the reserve of water decreased by 7.6×10^6 m³ (22%). The melted snow of the neve zone takes almost no part in forming the river flow. Infiltrating from the surface deep into the depths, the melted snow soaks not only the layer of snow deposited during the summer, but escapes into the lower layers of the neve. The zero mark of the temperature, the layer lying close to the surface during the period of intensive melting and the formation in the snow of numerous icy layers from 1 mm to 4-5 cm thick, also confirm this. Apparently, the process of compacting the snow in the Fedtchenko Glacier basin takes place mainly under the influence of the infiltration of melted snow and compression.

3.3 *Thermal balance of the surface*

A quantitative appraisal of the processes taking place in the glacier's most active period of life—the period of intensive snow melting—has been given, after calculating the components of the thermal balance of the glacier's surface.

In the neve zone, a zone of the glacier's mass accumulation, the credit part of the thermal balance consists of the radiation balance (B) and the turbulent influx of heat (q), the debit part of an influx of heat into the snow due to heat conductivity (P), a flow of heat spent on evaporation (eE) and a heat flow, C , spent on melting. The mean daily courses of the components and the thermal balance is given on fig. 2a. Turbulent influxes of heat and moisture were determined on the basis of average profiles of the temperature and specific moisture according to the method proposed by A. M. Obukhov, A. S. Monin and A. B. Kazanski, a current of heat into the snow due to the heat conductivity was determined with the aid of a thermo-transitometer and by the profiles of the temperature of the snow close to the surface. The significance of all the components of the thermal balance expressed in per cent with respect to $B + q$ for the whole period of intensive snow melting is as follows:

B	q	P	eE	C
52.8	47.2	0.6	60	39.4

(the value quantity, C , obtained in accordance with the method of the thermal balance and by snow measuring surveys, coincided) The matter "prefers" to evaporate", thanks to which the quantity of melted matter seems comparatively small. And the quantity of evaporated matter is not great, thanks to the fact that the heat evaporation, for the snow is almost 10 times greater than the heat of melting. This conclusion is in accord with our idea of the glacier's neve matter accumulated in this zone as a result of the glacier's flow, enters its lower part where after intensive melting it feeds the rivers flowing from the glacier.

An entirely different picture is to be observed in the middle and lower parts of the glacier. There is no evaporation from the glacier's surface, instead there is a condensation of the water vapor on the surface of the ice, thanks to the fact that the air flowing over the glacier in this region has a moisture close to saturation. Thus, the factor hindering melting is absent and the entire heat found on the glacier's surface is practically spent on melting.

The values making up the thermal balance, expressed in ratio to the sum $B + q + CE$ including heat, released in the condensation process of the water vapor is as follows:

B	q	eE	P	C
49.2	46.5	4.3	0	100

Measurements comprising the thermal balance in the valley of the Seldar River flowing from the Fedchenko Glacier, directly before the glacier's tongue, give an idea of the influence of the proximity of the glacier on the heating regime of the valley's surface (Fig. 2c). The mean daily course of the vertical turbulent flow of heat is directed from below upwards, i.e. a convection takes place, faintly expressed at night and strongly expressed in the daytime. This is explained by the presence here of the mountain valley circulation and the character of the surface. The equation of the thermal balance for the region of the station is as follows: $B = q + P + eE$ the values making up the thermal balance, expressed in per cent in ratio to the quantity of the radiation balance in per cent is:

B	q	P	eE
100	69	2	35

The observations conducted of the thermal balance of the glacier's surface, its movement, accumulation and melting of matter afford an idea of the role of the glacial feeding of rivers.

THE PRESENT PHASE OF INTRASECULAR VARIABILITY OF MOUNTAIN GLACIATION IN NORTHERN HEMISPHERE

A.V. SHNITNIKOV

Laboratory of Limnology USSR Academy of Sciences

SUMMARY

1. The problem of fluctuations, or movements, of mountain glaciers has been attracting the attention of scientists for a long time. The principal reason for it are those sudden catastrophes which they bring from time to time to the settlements of man situated near them. (Alps, Scandinavia, Iceland, Himalayas).

Within the last century, fluctuations of mountain glaciers of Europe and North America riveted the attention of scientists, hence a large number of important investigations into this problem.

It should be noted, however, that the vast majority of such investigations are concerned with the study of glaciers of individual mountain ranges or geographical zones. Meanwhile, comparison of conditions and character of the variability of mountain glaciers in various zones of the Northern Hemisphere is of great interest from the viewpoint of analysis of common developments in geographical phenomena.

2. At least two principal series should be distinguished in the fluctuations or movements, of mountain glaciers :

(a) fluctuations resulting from « multisecular » developments of the geographical sphere components, which are well pronounced throughout the Holocene;

(b) fluctuations consequent on « intrasecular » variations in climate of Brückner's type cycles, i. e. « intrasecular » variability of the condition of mountain glaciers in the Northern Hemisphere.

The first of these manifest themselves in the present epoch, beginning from the middle of the 19th century or, in certain areas, from the second half of the 18th century (Iceland, North America, possibly Scandinavia), in a consistent global retreat of all mountain glaciers. This phenomenon marks the beginning of a predominantly warm and dry phase of the last (VIIIth) stage, i. e. the Fernau stage, in the disintegration of the Würm (Valdai) glaciation.

The second are cycles of advance and retreat of mountain glaciers whose duration, although varying within a wide range, rarely exceeds 20 to 30 years. They occur against the background of multisecular variability, appearing as a sort of ripple on the long waves of the latter.

Both are characterized by two phases: a rapidly and vigorously developing transgression and a quiet and prolonged regression.

3. It is especially pertinent to emphasize this division of all principal movements of mountain glaciers in the present epoch since in world literature there is no single conception concerning it. It is for this reason that opinion has been expressed of the last general advance of mountain glaciers in the 17th-19th centuries as of a « minor ice age ». However, it was but the transgressive cool - and - humid phase of the VIIIth post - Würm multisecular stage, i. e. the maximum of glacial development in the Fernau stage.

This thesis has been elsewhere demonstrated by the author on the basis of comparing variations in the depression of the ice boundary in the Alps, Caucasus and Altai in the Helocene epoch with those in the movements of Fennoscandia in the same epoch and in the development of some other components of the geographical sphere. At present, an analysis of new data — variations in the altitudes of end moraines in the same epoch, as excellent natural indicators of their variability — conclusively corroborates this thesis, thereby confirming the difference between the series of multisecular and intrasecular fluctuations of mountain glaciers.

The division of all fluctuations of glaciers into two principal series is an extremely important factor in forming a correct conception of many contemporary general and specific processes and regularities of physiography and of present variations in its components. This primarily concerns changes in mountain glaciation, but is also true in respect of such regularities as, for instance, those of the total runoff in the Aral basin, of the general process of fluctuations in the Caspian level and so on.

4. At present our knowledge of fluctuations of mountain glaciers in the intrasecular variability rhythm (Brückner-type cycles) is more detailed with reference to the Alps as well as Iceland and, partly, Fennoscandia, these being areas of oldest settlement with glaciers playing a considerable role in the life of local population.

The first historically recorded stage of intrasecular variability coincided here with the beginning of a general advance of mountain glaciers throughout the Northern Hemisphere in the development of a cool and humid Fernau stage of multi-secular variability, dating back to the last years of the 16th century or the first years of the 17th century.

In the period of the 17th to 19th centuries, ten stages of intrasecular variability have developed here, one or two of them in the second half of the 19th century being related to the beginning of the general retreat of mountain glaciers of the world in the course of the development of the predominantly warm and dry phase of the multi-secular Fernau stage.

Thus, intrasecular variability of mountain glaciers in this particular case develops against the well defined background of their multi-secular variability.

5. The scantier knowledge of the glaciation of all other mountains in the Northern Hemisphere and the lack of historical records concerning them due to the relatively recent settlement in the respective regions make it impossible to analyse, with the same degree of duration and detail, the intrasecular variability of their glaciers. Nevertheless, it appears possible to determine the variability of mountain glaciers for some mountains of Eurasia and North America in the 19th century and, partly, in the second half of the 18th century. Analysis reveals a considerable degree of synchronism of individual intrasecular variability stages for all mountains of the Northern Hemisphere, expressed in a general form, in its principal manifestations. Thus for the Altai and the Caucasus, advance phases of the « twenties » and « fifties » of the 19th century are known, being well defined in the mountains of Europe and North America.

At the same time, in all mountains a considerable degree of asynchronism is to be noted in the movements of individual glaciers, due to their size, location, exposure, peculiarities of the feeding conditions, etc.

6. The least elucidated in world literature is the question of intrasecular variability of mountain glaciers in the first half of the present century.

Material available for various mountains of the Northern Hemisphere allows to establish that typical for a vast majority of them was an advance phase in the tens and early twenties of the present century.

At the same time there was a certain asynchronism, e. g. in the shifting of phases towards lead of the beginning of advance in Northern Europe, which, however, did not violate the general regularity of intrasecular variability.

7. In conformity with the general character of intrasecular variability of fluctuations of glaciers in time, following the phase of « tens » and « twenties », a phase of advance of glaciers should have been expected to commence in the middle or late forties and continue through part of the fifties of the present century.

However, from the late twenties and early thirties, under the influence of more general causes, the process of mountain glacier retreat became particularly progressive everywhere, continuing almost in all regions up to the present time (1959-1960).

8. Numerous investigations in recent years reveal regularities in fluctuations of glaciers as dependent on current variations in climate.

Intrasecular variability of the condition of glaciers is also quite definitely dependent on intrasecular fluctuations of climate, in particular, on long-term variations of air temperature in Northern Europe and on the general amount of ice in the North Atlantic. This circumstance aids to elucidate the condition which prevails at present (1959-1960).

Since the general climatic condition in Northern Europe (being a factor in and at the same time, an indicator of fluctuations of mountain glaciers) in the course of the forties favoured their expansion, an advance phase in their intrasecular variability should have been expected to commence in the late forties or early fifties.

9. From the late forties a considerable number of glaciers in Alaska showed a trend, sometimes quite intensive, towards advance. From the early fifties many of them began a well pronounced steady advance. Judging by the publications which have appeared so far, this trend has not manifested itself so clearly anywhere else. Moreover, glacier retreat is continuing on the vast majority of mountains in Eurasia. However, oftentimes a slower rate of retreat is to be observed, less frequently the retreat is halting, while sometimes a small advance is to be encountered. This phenomenon should be regarded as a manifestation of the current transgressive phase of the intrasecular variability of mountain glaciation in the Northern Hemisphere as a whole. In the glaciers of Alaska which more rapidly respond to variations in the environment, this phase is expressed in a full degree.

1. Dans les oscillations, ou mouvements des glaciers encaissés, il faut distinguer des oscillations appartenant au moins à deux « rangs » principaux :

(a) les oscillations en résultat de dynamique « multiséculaire » des composants du milieu géographique, exprimé à la perfection au cours de toute l'époque de l'holocène;

(b) les oscillations, représentant le résultat de variations « intraséculaires » du climat, selon le type des cycles de Brickner, c'est-à-dire une variabilité « intraséculaire » de l'état des glaciers encaissés de l'hémisphère nord.

La première d'entre elles, à commencer par le milieu du XIX^e siècle, ou ça et là dès le milieu du XVIII^e siècle (Islande, Amérique du Nord, peut-être la Scandinavie) est exprimée par un recul dirigé et global de tous les glaciers encaissés. Ce phénomène représente le début d'une phase principalement chaude et sèche du dernier (VIII) stade, c'est-à-dire la phase fernaou de la désagrégation de la glaciation wurmienne (Valdai).

La deuxième — représente des cycles d'approches et de reculs des glaciers avec une durée des cycles, qui varie en grandes limites, mais qui, à la moyenne, dépasse rarement 20-30 années. Elle a lieu sur le fond d'une variabilité multiséculaire et représente une sorte de ride sur les longues ondes de cette dernière.

Pour les deux variabilités caractéristique est une division en deux phases : phase transgressive se développant rapidement et impétueusement et phase régressive, calme et de longue durée.

2. Une telle division de tous les principaux mouvements actuels des glaciers encaissés doit être particulièrement soulignée, car il n'y a point de notion unique sur cette division dans la littérature mondiale. C'est précisément pour cette même raison que l'on expose parfois des points de vue sur la dernière approche générale des glaciers encaissés, qui eut lieu aux XVII-XIX^e siècles, comme sur un « petit siècle de glaciation ». Cependant, ceci était seulement la phase transgressive, froide-humide du stade post-wurmien multiséculaire, c'est-à-dire le développement maximum de la glaciation au cours du stade fernaou.

3. La notion la plus distincte sur les oscillations des glaciers encaissés dans le rythme de leur variabilité intraséculaire (du type des cycles de Brickner) existe actuellement en rapport aux Alpes, de même qu'à l'Islande et en partie à la Fennoscandie.

Le premier stade de variabilité intraséculaire, mentionné dans les notes historiques, coïncida ici au début de l'approche générale des glaciers encaissés sur tout l'hémisphère du nord dans le développement de la phase froide-humide du stade fernaou de la variabilité multiséculaire et se rapporte aux dernières années du XVI^e siècle ou aux premières années du XVII^e siècle. Au cours des XVII-XIX^e siècles se réalisèrent ici 10 stades de variabilité intraséculaire; l'une ou deux d'entre elles, appartenant à la deuxième moitié du XIX^e siècle, font déjà partie du recul général des glaciers encaissés du globe terrestre au cours du processus de développement, principalement, de la phase chaude et sèche du stade multiséculaire de fernaou.

4. L'étude générale moins détaillée de la glaciation de toutes les autres structures montagneuses de l'hémisphère nord et l'absence de notes historiques sur elles ne donnent point la possibilité d'analyser avec le même degré de détails la variabilité intraséculaire de leurs glaciers. Néanmoins, l'on peut établir la variabilité au cours du XIX^e siècle et, en partie de la deuxième moitié du XVIII^e siècle pour une série de constructions montagneuses de l'Eurasie et de l'Amérique du Nord. L'analyse exécutée indique un synchronisme considérable des stades séparés de la variabilité intraséculaire pour toutes les constructions montagneuses de l'hémisphère nord, exprimé en forme générale dans ses manifestations principales. Ainsi, par exemple, (pour l'Altai et le Caucase sont connues les phases d'approche des années « 20 » et « 50 » du XIX^e siècle, bien et clairement exprimés, dans les structures montagneuses de l'Europe et de l'Amérique du Nord.

En même temps, dans toutes les constructions montagneuses se manifeste un synchronisme considérable des mouvements des glaciers séparés, en fonction de leurs dimensions, de leur situation, exposition, des particularités dans les conditions d'alimentation, etc.

5. La moins éclairée en littérature mondiale reste la variabilité intraséculaire des glaciers encaissés au cours de la première moitié du siècle courant.

Le matériel, accumulé sur certaines constructions montagneuses de l'hémisphère nord, permet d'établir que pour leur majorité prépondérante était typique la phase d'approche au cours des années 10 et durant la première moitié des années 20 du siècle courant.

Un certain asynchronisme eut aussi lieu; il s'exprimait, par exemple, pour le nord de l'Europe en un déplacement des phases vers un devancement du début de l'approche, qui, néanmoins, ne violait point la loi générale de la variabilité intra-séculaire.

6. En accord au caractère général de la variabilité intraséculaire des oscillations des glaciers en temps — l'on devait s'attendre, après la phase des années 10-20-30 à une phase d'approche des glaciers (de la moitié ou de la fin des années 40 et au cours d'une partie des années 50 du siècle courant).

Néanmoins, dès la fin des années 20 et du début des années 30, le processus du recul des glaciers encaissés, sous l'action de causes à caractère d'ordre plus général, a acquis universellement un progrès particulièrement fort. Presque universellement il continue aussi actuellement (1959-1960).

7. Les nombreuses investigations, au cours des années récentes, révèlent les lois des oscillations des glaciers en fonction des variations courantes du climat.

Et la variabilité intraséculaire de l'état des glaciers se trouve aussi en fonction bien exprimée des variations intraséculaires du climat. Notamment, l'on trace une dépendance bien exprimée des variations de la température de l'air dans l'Europe Septentrionale entre la marche de longue durée et la glaciabilité générale de l'Atlantique Septentrionale. Cette circonstance aide à expliquer l'état qui se créa vers le temps actuel (1959-1960).

8. Dès la deuxième moitié des années 40 pour une grande quantité des glaciers d'Alaska se manifesta une tendance à une avance parfois très intense. Dès le début des années 50, beaucoup d'entre eux passèrent à une approche stable, clairement exprimée. En tant que l'on peut éclaircir de la littérature accumulée jusqu'à présent, cette tendance n'est nulle part ailleurs exprimée si clairement. Bien plus, sur une majorité prépondérante des glaciers des constructions montagneuses de l'Eurasie, le recul continue.

Occasional large-scale movements of mountain glaciers have since long attracted the attention of people. The primary reason for it are those sudden disasters which they from time to time bring to the settlements of man located near glaciers. This is particularly true in respect of long-settled and rather densely populated regions of such regions where the movements of mountain glaciers greatly affect the invariable conditions of life of the population, including the afore-mentioned disastrous consequences. The former regions in the first place include the Alps in all their diversity. The latter are represented by Iceland, Scandinavia, many regions in Tien Shan, the Pamirs, the Himalayas, etc.

The historical chronicles of Switzerland, the Alpine part of France, Austria, Iceland, and of the Scandinavian states for the last few centuries abound in reports of sudden and disastrous advances of mountain glaciers on mountain settlements.

In the last century, fluctuations of glaciers of the mountains of Europe and North America attracted the attention of scientists, which already at the close of the 19th century resulted in the appearance of a series of studies dedicated to this important problem in Europe (Heim, 1887; Richter, 1891; Rabot, 1897; Forel, 1900, and others). The last in this list of authors is known to have set up in the Alps in the eighties of last century a special service for studying fluctuations of glaciers which has been successfully operating up to the present time. Later on, due to the great importance of this phenomenon for many aspects of the economic activities of nations, the study of fluctuations of mountain glaciers assumed a world-wide scope. Hence, a series of new summaries covering individual glacier regions (Kalesnik, 1937; Kalesnik, 1938; Matthes, 1942; Thorarinnsson, 1943; Lawrence and Elson, 1953; Tollner, 1955 and many others) or dealing with the history of fluctuations of individual glaciers (Mercanton, 1916; Mouglin 1908 and 1926, and others). At present there are a great many publications concerned with the fluctuations of mountain glaciers of the world, particularly in respect of the glaciers of Europe. The situation is far worse as regards the data on the movements of glaciers in the mountains of the USSR, North and particularly, South Americas, and South-East Asia. However, we shall not dwell here on this problem at all. It only seems relevant to point out that, as far as we know, with the exception of S. V. Kalesnik's monograph "General Glaciology" (1939) up to the present time there are no studies which would summarize the phenomena of glacier fluctuations for whole continents or hemispheres. In the meantime, such

summary is now quite essential from the point of view of the possibility of forecasting, although in very general terms, the further trend in the fluctuations of the world's mountain glaciers in time.

Before proceeding to the analysis of the data on the present-day movements of mountain glaciers, it seems necessary to briefly outline their most important regularities in the post-glacial time.

Phenomena of at least two principal "ranks" should be distinguished in the fluctuations of mountain glaciers:

a) fluctuations consequent on the "multisecular" changes of the geographical sphere components well defined throughout the post-Würm epoch, i.e. during the last 13000-15000 years;

b) fluctuations resulting from "intrasecular" variations in climate of the type of Brückner's cycles, i.e. "intrasecular" variability of the condition of mountain glaciers.

There may be fluctuations of some other "ranks", e.g. "secular" fluctuations. Up to the present time, however, "secular" or any other variability in the movements of mountain glaciers has not been actually established anywhere, although on the basis of analysing the movements of Bossonf glacier Mougín (1926) surmises the existence of "secular" variability.

As to the two afore-mentioned regularities in the movements of mountain glaciers, the first of them—the multisecular—consists in rhythmic fluctuations, the duration of individual rhythms being somewhat less than 2000 years (author, 1949, 1953, 1957 and others). Each rhythm, being a stage in the disintegration of the Würm glaciation, is divided into two phases: the short and intense advance phase of 250 to 350 years, and the slow and quietly developing retreat phase lasting for about 1000 years. Seven such rhythms or stages have passed since the beginning of disintegration of the last glaciation. This regularity for some mountains of Eurasia is shown in Fig. 1.

In order to obtain this diagram, in accordance with the general regularity previously indicated by us (1953 and 1957), use was made of the data on the altitudes of stage moraines given by N.N. Palgov for the Trans-Iliy Alatau, by P.A. Cherkasov for the Dzungari Alatau, by I.M. Miagkov for the Altai (Katoun), by S.L. Koushev for the Caucasus (Benzengi). Shown in the diagram are the altitudes of stage moraines, beginning either from the maximum Würm moraines, whenever their altitudes are known, or from the first stage of retreat of the last glaciation (Shlieren—after the Alpine scheme) almost all of which are known. It may be pointed out that the altitudes of some yet undiscovered moraines of the Trans-Iliy Alatau were predicted by us on the basis of the afore-mentioned regularity (1953); one of such moraines was afterwards actually discovered by P.A. Cherkasov (1957) at the height expected by us.

Dotted curves between the points of height position of stage moraines indicate the direction of the movement of glaciers in the retreat phases of the respective stages. In each stage the position of the ends of glaciers at the time of their transition from the retreat phase to the advance phase is not preserved in nature and thus remains unknown; therefore these transitions are shown as broken in the diagram.

The diagram represented in Fig. 1 shows how strict is the regularity of the multisecular variability of mountain glaciation in the course of the disintegration of the Würm. Stage moraines, these excellent natural indicators of the phenomena which took place thousands or many hundreds years ago not only show the sequence of the processes of glacier movements in each of the mountain systems represented here, but also indicate the individual peculiarities of each of them. Depending on the physiographical peculiarities, each of the mountain system has certain altitude limits of ancient and present-day glaciers, as seen from the table below:

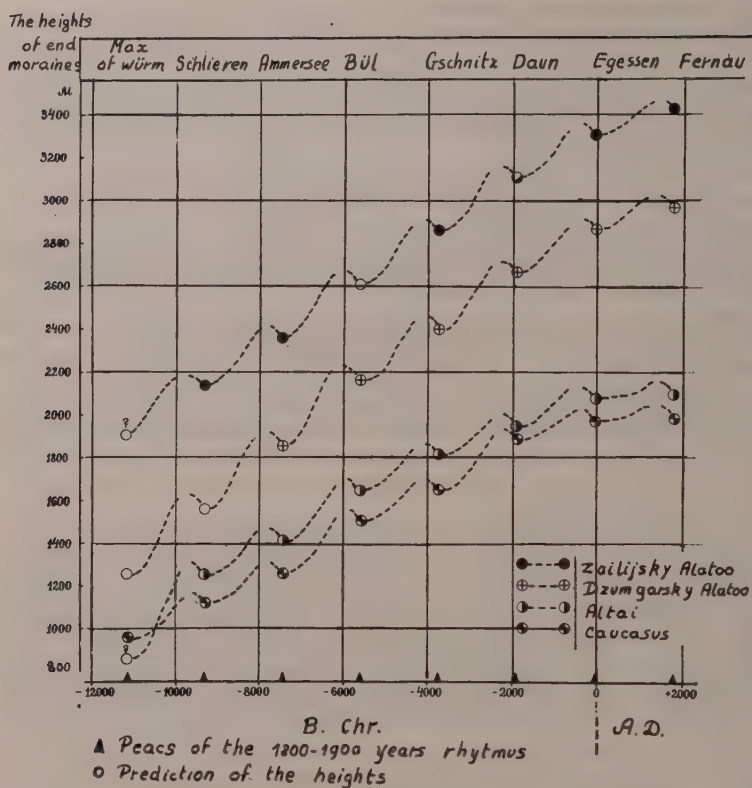


Fig. 1 — Heights of the end moraines of some mountains of Eurasia on the stadials of postwürm (postvaldai).

Mountain system	Altitude of stage moraines	
	Shlieren	Present
Trans-Iliy Alatau	m 2130	m 3425
Dzhungari an Alatau	1550 (?)	2966
Altai	1250	2090
Caucasus	1110	1980

It follows from the above, as well as from Fig. 1 that our times, in the course of the disintegration of the last glaciation, as regards mountain glaciers belong to the beginning of the eighth post-Würm stage. It was preceded by the seventh stage, Fernau.

tage (by the Alpine scheme) which ended by a global advance of mountain glaciers in the period of 17th-19th centuries, sometimes referred to as the "17th-19th centuries stage".

The contemporary beginning of the eighth post-Würm stage is at the same time the beginning of its retreat phase of predominantly warm and dry phase in respect of the general moisture supply of continents (author, 1957). As regards glaciers it manifests itself in a directional global retreat of all mountain glaciers which commenced in the middle of the 19th century, and in some areas in the second half of the 18th century (Iceland, possibly North America? and Scandinavia).

The second of the two afore-mentioned regularities of glacier movements—the intrasecular regularity—consists in perpetually recurring cycles of glacier advances and retreats which, although varying within considerable time limits, are on the average far beyond 25 to 35 year periods. Intrasecular variations of mountain glacier condition occur against the background of multisecular variability, being a sort of ripples on the long waves of the latter. The relationship between multisecular and intrasecular variability is shown in Fig. 2 which is self-explanatory. Like multisecular fluctuations, intrasecular variability is characterized by a well pronounced division into two phases: a rapidly and vigorously developing transgressive phase with glacier advances, and a quiet and relatively long regressive phase with glacier retreats. It is caused by climatic fluctuations of the Brückner-cycle type which constitute the subject matter of numerous studies.

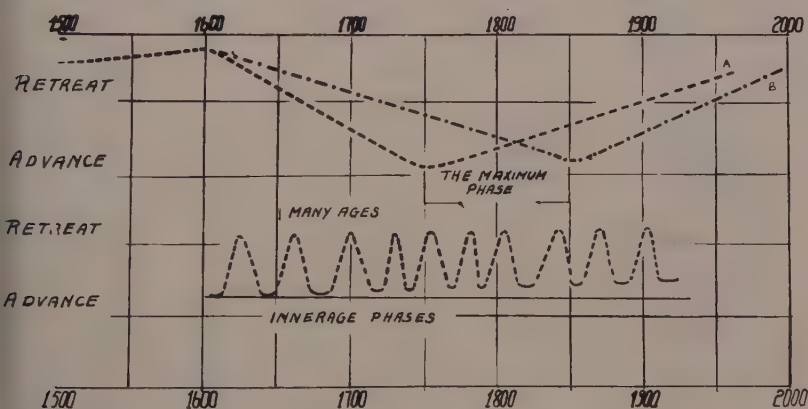


Fig. 2 — Connection between the many ages (1800-1900 years) and the innerages changes of the mountain glaciation (the Alps as the example).

It seems particularly relevant to emphasize the division of the principal types of contemporary mountain glacier movements as there is no common conception of it in world literature. It is for this reason that opinion is oftentimes expressed of the last global advance of mountain glaciers in the 17th-19th centuries as of a "minor ice age" (Ahlmann, 1953; Erinc, 1952; Lawrence and Elson, 1953; Holms and Moss, 1955; Sharp and others, 1959). Meanwhile it is implied that in the period between the maximum stage of the last glaciation and the time preceeding the beginning of glacier advance in the "17th-19th centuries" stage, i.e. in the 14th-16th centuries, they never reached that degree of expansion which they attained at their maximum in the 18th-19th centuries. This opinion is quite erroneous. That was but a transgressive, cool and humid phase VII of the last post-Würm multisecular stage, i.e. the maximum of glaciation development in the Fernau stage. Moreover, it was in the

TABLE 1

Fluctuations of glaciers of the Alps 17th-18th centuries

	Richter 1891 East. Alps	Rabot 1914-15 Chamonix Valley	Mougin 1908-1926 Bossons Glacier	Klebelberg 1948 Grindel- wald Glacier 2 & others	Rabot 1897 North Norway	Rabot 1897 Iceland	Averaged approx. years	Intervals, years
1600	ab. 1600-s	1609-11	1605-10	ab. 1595	1600-02		1600-1610 (1605)	
	1630-40	1641-44 1664	1643	1620	1632-34		1632-44 (1638)	33
1650								
	1675		1685		1685-87 1695-97		1664-85(?) (1675?)	37
1700	1712-15	1715-16	1712(?)				1712-16 (1714)	39
	1735-40	(?)	?	1719		ab. 1734	1734-43 (1738)	24
1750	1767-70	ab. 1770		1743	1741-43	ab. 1751		
			1787	1768 1770 1779		1783-(88)	1767-70 (1768)	30

	Forel 1900 Austr. & Swiss Alps	Heim 1885 (Alps)	Richter 1891 (Alps)	Rabot 1914-15 Klebel- berg 1948 (Chamo- nix Valley glaciers)	Mougin 1908- 1928 (Bossons Glacier)	Brück- ner 1909-10	Hey- brock 1940 (Brenva)	Guet 1929 (Trient & other glaciers)	Tollner (Eastern Alps) 1954	Klebel- berg 1948 Jost 1958 Grindel- wald glacier	Oech- slin (1951 & (1958)	Averaged years	Intervals years
1750			1767-70	advance abt. 1770-s					max. abt. 1750-s adv. abt. 1770-s	abt. 1743 1768 ; 1770 ; 1779		1767-70 (1768)	
1800		1811-22	1814-20	1818-25	1787 1818 max.	between 1815-25	1818 max.	1820 max.	adv. abt. 1790-s inten- sive adv. abt. 1820-s	1818-22		1787-90 (1789) 1814-25 (1820)	21 31
1850	1811-20 (1818-20 gr. max)	1840-50	1840-50	1850-55	1850-54 max.	1845-55 partly to 1860 and on 65	1850 max.	1850 max.	princi- pal max. 19th cent. by 1856	1844-56		1845-56 (1852)	32
	1876-92 Sw. 1890- 900 Austr.	1880		1878-94	1880-97 1889 1892 & 1896	1890-95	1897 max.	1878-95	Cessa- tion of retr. and adv. abt. 1890 and abt. 1900	late 1890-s general cess. in retr. abt. 1890		1880-94 (1888)	35
1900					1921 max.		1915-23	1916-24	1910-20	1913-24	1910-25	1914-24 (1919)	31
1950							1938 max.						

smallest stage, too, as with each succeeding stage the expansion of glaciers diminishes due to the very nature of the phenomenon of disintegration of the Würm glaciation, the glaciers becoming increasingly smaller in their retreat phases as well.

This has been shown by the author (1949, 1953, 1957) on the basis of comparing the changes in the snow-line depressions of the Alps, the Caucasus and the Altai in the post-Würm period with the changes in the vertical movements of the Fennoscandian massif in the same epoch, as well as with the development of some other components of the geographical sphere. Now, on the basis of the analysis of new material—changes of end moraines altitudes in the same epoch, the length of the glaciers themselves in various stages, as excellent indicators of their variability, this thesis has been further corroborated.

On the basis of the available summary works mentioned above and the works dealing with analysis of variations of individual glaciers from the close of the 16th century to the present time, we have compiled comparative tables of the movements of glaciers in the Alps and the North Atlantic in the same period (Tables 1 and 2), i.e. prior to the beginning of the 20th century, as regards their advance phases.

Without dwelling on these tables, it should be noted in the first place that the dates of glacier advances in various regions of Europe given on the basis of widely diverse sources are very close to each other and usually do not vary within more than 6 to 10 years. This variance is quite natural, depending partly on the geographical (zonal) position of glaciers and partly on widely diverse local conditions.

It is important, however, that all the dates make up quite definite groups which are easily averaged as has been done in the two tables. It appears that from the turn of the 17th century when mountain glaciers began large-scale advances in connection with the beginning of a regular cool and humid phase of multiseccular variability, as outlined above, and up to the beginning of the current century, i.e. during 300 odd years, the glaciers of the Alps and Northern Europe went through 10 stages of intraseccular variability. As seen from Tables 1 and 2, the time intervals between the phases vary within 21 to 29 years. It should be noted that out of the 10 stages only two lasted for 21 and 24 years, and one for 39 years; the duration of all the other, i.e. 70 percent, varies within 30 to 37 years, the principal amplitude of variations thus being only 7 years (Fig. 3).

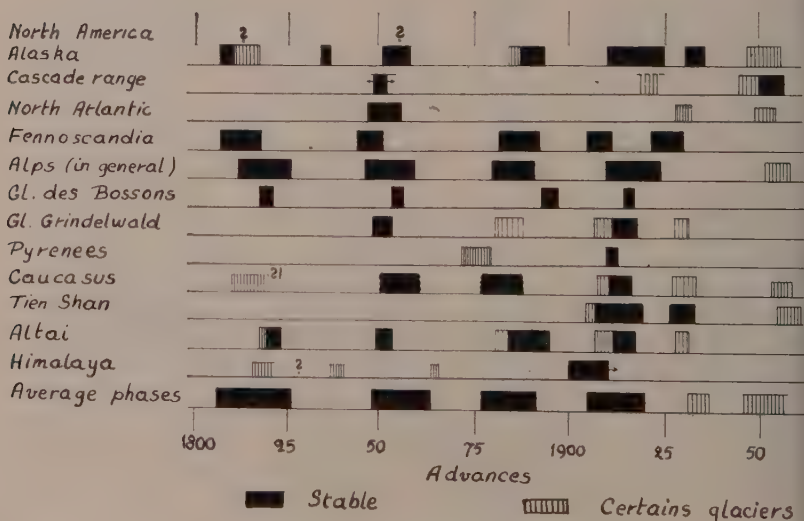


Fig. 3 — Innerage changes of the glaciers.

The data on the movements of glaciers in the Pyrenees are, unfortunately, extremely scarce. However, from Barrere's data it appears that here, against the background of the general retreat commencing in the middle of last century, there was a well pronounced advance of glaciers in the middle seventies and, possibly, in the eighties of the 19th century, and then in the tens of the current century. Thus, the last phases of the advance of these glaciers appear to be synchronous with those of the Alps.

The comparisons made indicate that for the mountain glaciers of Europe as a whole and for the North Atlantic the multiseccular variability of their condition is a stable phenomenon regularly recurring during more than three centuries. A well defined feature is the basic synchronism of some movements both of individual glaciers, depending on their peculiarities, and, possibly, of the glaciers of whole regions, depending on their geographical zonality. To a certain extent it concerns the glaciers of Iceland, Spitsbergen and partly Norway, where some phases are somewhat advanced in time, which is not only a quite natural phenomenon but also a very interesting and important geographical fact. However, we shall not dwell on it since we are now concerned with general regularities.

Material relating to intraseccular movements of glaciers in the mountains of the USSR: the Caucasus, Tien Shan, Altai and Northern Urals cannot as yet be collected for as long a period as for the afore-mentioned areas, although for the Caucasus and Tien Shan corresponding investigations have already been undertaken. For the time being, it is possible to establish for them the principal stages of movements only for the 19th and the beginning of the 20th century.

The data obtained on the basis of analysing the works of many authors (Abikh, Zieger, Rossikov, Moushketov, Preobrazhensky, Kalesnik, Pavlov, Tronov, Toushinsky and many others) are outlined here in a general form.

It will be seen that well defined for the Caucasus were glacier advances in the middle and the last third of the 19th century (about 1880-87). They were followed by the advance at the end of the first decade and in the second decade of the current century.

Unfortunately, it has been impossible so far to find how well pronounced was in the Caucasus the advance of glaciers in the first third of the 19th century, about 1820-s, which was so manifest in all the mountains to the west of it. However, as a result of some recent investigations, by S.L. Koushev in particular, there is no longer any doubt as to its having taken place.

Thanks to the thorough investigations by M.V. and V.V. Tronovs, the phases of glacier advance in the 19th century in the Altai have been much better determined. Glacier advances were extremely well marked here about 1820-s, 1850-s and in the eighties of the 19th century. The first two advances are well confirmed by moraines designated as the "1820-s stage" and the "1850-s stage"; the 1890-s stage is also confirmed by moraines. Afterwards there was an advance in the period between 1911 and 1932 of different glaciers in different years, which, however, acquired the most general character in about 1917-18.

For Tien Shan (both Western and Eastern) the available data appear to be most scarce. Distinctly determined is the glacier advance of 1914-1924 and, partly, an apparently earlier advance (1907-08), while in some places a later advance up to 1931-32 depending on the local conditions.

There are indirect indications that an advance occurred here in 1880-s-1890-s. A comparison of all available data permits to state that glacier advances must inevitably have taken place in the first third and in the middle of the 19th century in Tien Shan as well.

To avoid possible misunderstanding it should be pointed out again that we are dealing with certain general, averaged phenomena in the movements of glaciers.

Some of them show rather considerable divergencies from the dates indicated here. In some cases, glaciers do not appear to undergo appreciable variations in the direction of their movement, for instance the Fedchanko glacier whose retreat, first noted in the seventies of last century, according to some observations has not yet stopped. However, it may be stated that this does not mean that during the period of almost one hundred years there has been no cessation or deceleration of retreat in its regime in the advance phases of the eighties of last century and the first two decades of the current century; such decelerations can occur with a delay against advance phases due to its particularly great dimensions. Conversely, small glaciers especially characteristic for the North (Scandinavia, Iceland), undergo changes in movement in time periods somewhat shorter than those for the medium-sized glaciers.

It is still more difficult to present a more or less clear picture of the glacier movements in the mountains comprising the mountain massif of South Asia (Kueng-Loun, Kara Korum, Himalayas). It is only from the material of Visser's expedition (1928) that some data can be obtained. Thus, among others, he considers with more or less detail the movements of Koomdan glacier. The road passing at its basis is covered by it in the periods of its advance and then cleared again. The condition of the road in the 19th century will be seen from the following table:

Cleared (i.e. the glacier not in the condition of maximum)	Covered (i.e. the glacier in a condition close to maximum)
1812 1865 1889-90; 1898 1909-1911	from 1818 to (at least) 1840 1869 1903; 1908

In addition to these data, Visser states that as he found from different sources the glacier:

Retreated:	Advanced:
about the beginning of 19th cent. from 1860-70-s to at least 1902 with some fluctuations	about 1700-s from 1825 to 1860-70 with many minor retreats between 1830 and 1842 from 1902 to at least 1911 (no further data available)

The data of systematic observations of the movement of 20-24 glaciers conducted from 1900 show that in the period of 1900 to 1910 out of the 21 glaciers under observation 18 (i.e. 86%) were advancing, while only 3 were retreating. In the period of 1911 to 1920 out of the 24 glaciers (of which at least 17 belonged to the first group as well) only 6 were advancing (25%), while 14 were retreating and 4 were stationary. Later, according to the observations of the expedition, the majority of the glaciers were apparently in the stationary condition. Thus, as regards the glaciers in a considerable section of the South-Asian massif, it may be supposed that the majority of

them were advancing in the first decade, and retreating in the second; in relation to all previously considered groups of glaciers a kind of delay, i.e. a shift of phases, is to be observed here.

The data on the fluctuations of glaciers in North America in the 19th century and earlier, available in literature, are scarce. This is not surprising as all of them are located either in the territories far to the north which are scarcely populated even nowadays, or in mountain regions seldom, if ever, visited in those times. However, from the works of Lawrence and Elson, Harrison (1956-57), Matthes (1946), Meier and others the following conception may be formed as to the movements of glaciers in Alaska, the Rocky Mountains and some other mountains in the U.S.A. Thus, Lawrence and Elson believe that the glaciers of North America attained their greatest development in the middle of the 18th century, after which a general retreat followed which has been continuing up to the present time. However, this retreat was repeatedly interrupted with ensuing short advances lasting for some years. Among such phases those of about 1744 and 1766 are noted as well as about 1835, 1861 and 1882. Besides, according to other data (Meier and others) an advance phase has been determined for the first decade of the 19th century, the 1880-s phase lasting to the early nineties. According to some data, well pronounced for the glaciers of the state of Washington (Harrison) is the phase of glacier retreat from the maximum in 1857 up to 1871 and, possibly, afterwards. Thus, in the movement of North American glaciers there has also been the advance phase of the middle of the 19th century which took place everywhere, this being a very important factor. Finally, a well pronounced advance phase is that in the tens of the current century which may have lasted to the middle twenties.

Comparison of the afore-mentioned advance phases of North American glaciers with the corresponding movements of Alpine glaciers shows that they have common features and are marked by a considerable degree of synchronism. Exceptional is the 1835 phase in which the Alpine glaciers were undoubtedly retreating. Possibility is not excluded that, in effect, it took place somewhat earlier, especially as the material given by Lawrence and Elson omits one of the most widely spread phases in the tens of the 19th century, this being hardly possible due to the fact that it was particularly well pronounced in all the mountains of the Northern Hemisphere.

The latter fact is confirmed by the material now available which indicates that the glaciers of Alaska and the Rocky Mountains in the 1910-s not only showed a trend towards advance but actually did advance noticeably (Rabot, 1913; Matthes, 1942; Lawrence and Elson; Bender and Haines). Heusser even points out that the Blue glacier deposited a well marked moraine in the second decade of past century. Dightman and Beatty give interesting climatological data on the glacier region in Montana, which provide a good basis for explaining the causes of the glacier advance in the Rocky Mountains and Alaska in the tens of the current century.

It is known that in Central Africa there is a number of glaciers on several peaks of Ruwenzori Range in the Kilimanjaro group. Although the glaciers of the mountains of Kenia, Stanley and others have been rather well explored, there are almost no data on the changes in their condition in the 19th century and especially in earlier periods. The only indication is Heinzelin's statement that prior to 1890 glacier tongues on Mountain Stanley underwent an advance phase after a preceeding retreat; this advance was followed by another retreat which intensified considerably in the latest decades. This single phase coincides in time with what we now know of the Alpine, North American and other glaciers.

Of great interest are some data on the movements of the glaciers of New Zealand given in Harrington's short summary. They were in the phase of greatest extension in about the middle of the 18th century (possibly in the forties) after which a retreat commenced which has been going on up to the present time with some interruptions during which the glaciers either remained stationary or slightly advanced. One of the

advance phases occurred in the middle of the 19th century as it was at about this time that the snow line began to rise in the mountains of New Zealand, and a retreat of the Rangitata group of glaciers and then, evidently, of some others commenced. Afterwards, in about 1890-s an advance phase came on again to be well marked by the "1890 moraine" and followed by the continuation of the general retreat of glaciers. As evidenced by the data of Harrington and others, there was undoubtedly one more advance phase in about the tens of the current century, in particular as regards Tasmann and Murchison Glaciers.

In the final analysis it may be supposed that the phase of the advance of mountain glaciers in the second decade was on the whole quite distinct for all the mountains of the Northern Hemisphere. In some areas it commenced in the first decade, and was oftentimes over by the first half of the third decade. Its greatest extension was to be observed in the second decade.

From the second half of the third decade, a regular phase of the general retreat of glaciers began to develop, becoming particularly well distinct in the second half of the fourth decade and in the first half of the fifth. It should be pointed out that after the thirties in some areas there was a check in their retreat or even a slight advance, however not on a large scale.

On the basis of the cyclic character of intrasecular movements of mountain glaciers and the average duration of individual cycles as stated above, i.e. close to 30 to 35 years, the next phase of their advance could be expected to begin in the late forties or early fifties.

Indeed, in some mountain regions of the Northern Hemisphere this trend manifested itself very distinctly. This first of all concerns Alaska and the territory of the Rocky Mountains in America (Hoffmann, Long, Bengtson, Harrison, Dightman and Beatty, Dyson and many others). Afterwards this phase manifested itself on the glaciers of other mountain regions as well: Spitsbergen (Mellor, Sweeting), Greenland (Wiedick), Sweden (Bergström), Iceland (Thorarinsson, 1956), etc. In the early fifties the percentage of advancing or stationary glaciers in the Alps somewhat increased (Fig. 4, also Finsterwalder, 1958, 1959). However, the general picture here throughout the whole decade, up to 1958, remains very motley and indistinct. Whereas in the central massif—the Swiss Alps—the number of advancing or stationary glaciers increased noticeably, in the Eastern Alps it changed but little, and in the Italian Alps this phenomenon practically did not manifest itself at all.

For the time being the data on the glaciers of the Caucasus, Tien Shan and Altai are scarce. However, as regards these areas there are also quite definite indications that in recent years there have been signs of some glaciers stopping or even slightly advancing against the background of the general rapid retreat. At the same time there are reports about the advance of some glaciers in the Nan Shan range in West China (Dolgoushin).

The foregoing allows a conclusion to be made that in principle the regular transgressive phase of the intrasecular variability of mountain glaciers, the phase of their advance, has been or is still developing on the whole during the fifties of the current century, i.e. in very recent years. In some areas it began in the second half of the forties, while its main part occurred in the fifties. On the basis of the usual duration of this phase, i.e. about 8 to 12 years, it may be supposed that the glaciers have already passed their principal development and in the next few years will enter the phase of their further retreat.

The weakly manifested character of the transgressive phase of the fifties is due to the fact that it is a component part of a multisecular phase of the general retreat of all mountain glaciers of the world in the course of the continuing disintegration of the last, Würm (Valdai) glaciation. The more intensively marked will be the intrasecular phase of glacier retreat which is setting in now. Thus the views outlined above

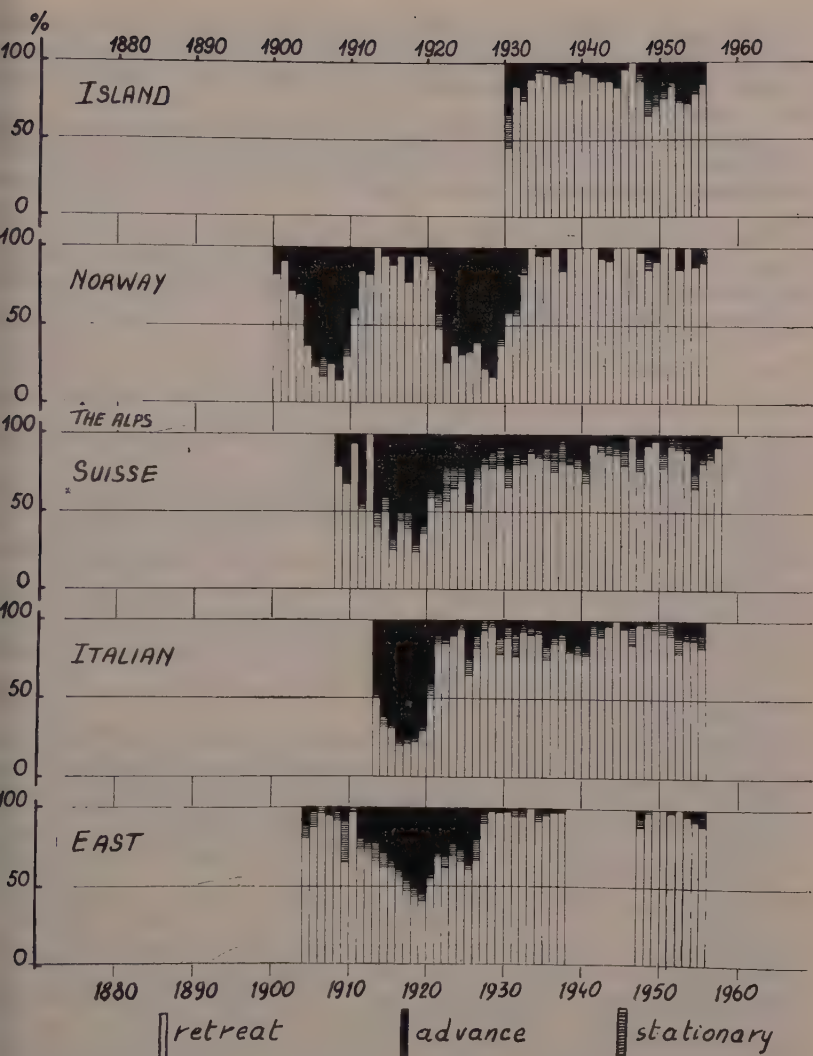


Fig. 4 — Innerages changes of glaciers.

re fully divergent from those of Prof. Vanni (1957) recently expounded by him in paper at the Hydrological Symposium in Toronto. Prof. Vanni believes that in the next few centuries and, consequently, in the next few years the glaciation of the Alps will gradually increase. It does not appear possible to set forth here his views on his most important problem. However, it follows from our conception, as outlined above, that:

1. Mountain glaciers of the whole world, in the course of the development of the regular (VIII) post-Würm multiseccular stage, in its regressive phase, will be diminishing during many centuries.

2. The glaciers, in the course of their intrasecular variability, in the next few years will enter a regular intrasecular regressive phase, i.e. retreat phase, which will replace the weakly expressed or nearly unexpressed transgressive phase, i.e. advance phase, which is ending now.

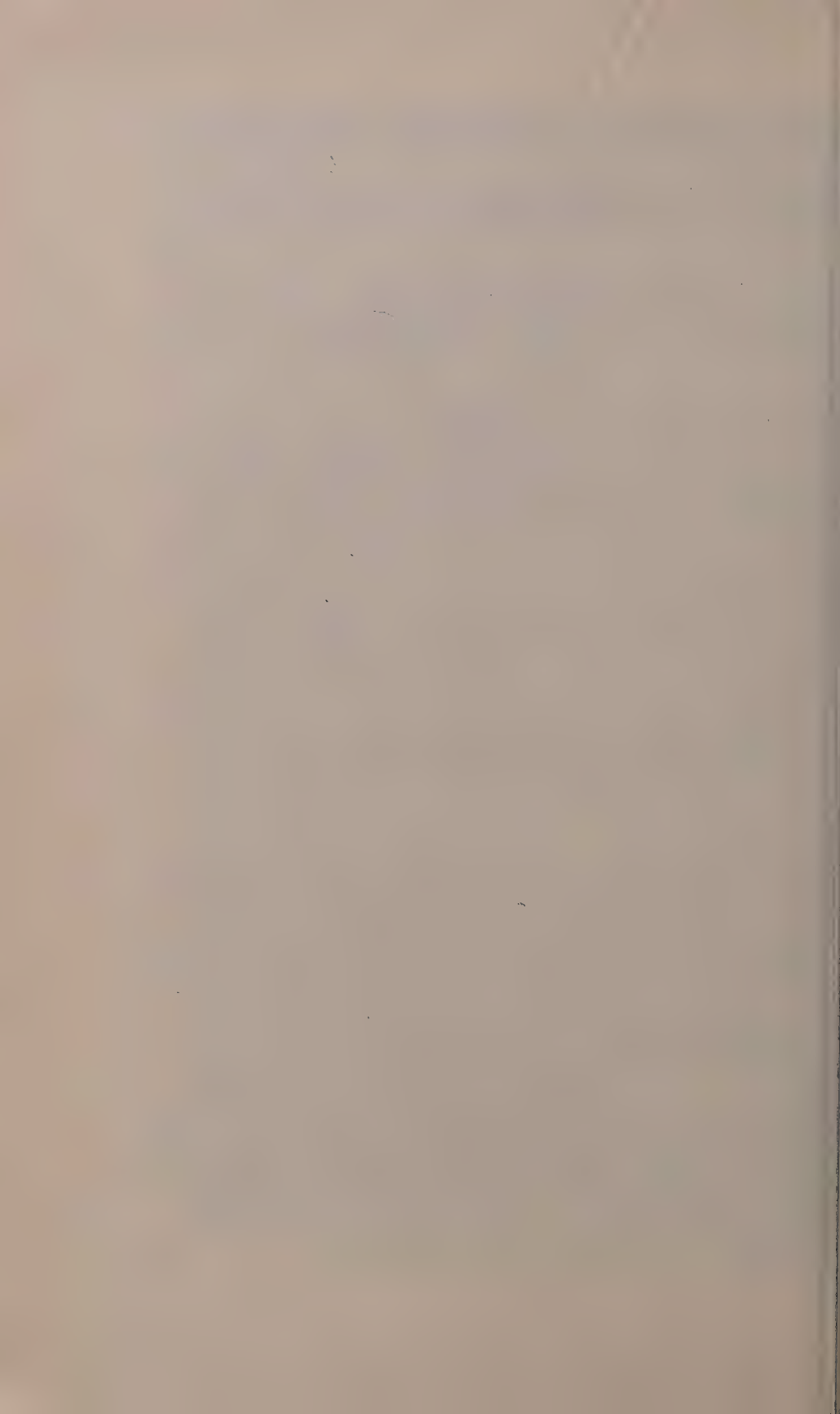
3. Inasmuch as the forthcoming regressive intrasecular phase is superimposed on the multisecular regressive phase, the retreat of glaciers in the next decade or two will be even more vigorously marked than in the thirties and forties of the current century.

4. This process may be somewhat weakened only if the abatement of solar activity, to begin in the next few years in the course of transition of its secular cycle, now at maximum, to a phase of intensive decrease, determines the appearance of some climatic phenomena non-existent throughout last century.

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MESURE DES GLACIERS

GLACIER SURVEYING AND THICKNESS MEASUREMENT

TELLUROMETER MEASUREMENTS ON THE GREENLAND ICE CAP DURING THE INTERNATIONAL GLACIOLOGICAL GREENLAND EXPEDITION (EGIG) SUMMER 1959

Walther HOFMANN

SUMMARY

During the summer campaign 1959 of the International Glaciological Greenland Expedition, profiles of fixed points on the Ice Cap were established by distance measurements with the Tellurometer. The physical and climatic conditions on the ice shield proved to be especially favourable for the Tellurometer measurements. In both the West-East-Profile from coast to coast and the North-South-Profile the reached accuracy for a measured distance is ± 4 cm. Glaciological results can be derived from these measurements only after their repetition, which is planned at the earliest in 4-5, at the latest in 8-10 years. Already in 1959, the differential movement of a section of the West-East-Profile near the West Coast was determined by repeated measurement. The section of 35 km was effected by a total expansion of 9.10 m in 3 months. — The Tellurometer has proved its aptitude for the establishment of fixed points on ice shields with high precision.

RÉSUMÉ

Durant la campagne d'été 1959 de l'Expédition Glaciologique au Groenland, on a installé des profils de points fixés à l'Inlandsis par mesure des distances avec le Telluromètre. Les conditions physiques et climatiques à l'Inlandsis se montraient particulièrement favorables pour les mesures tellurométriques. On a atteint dans les deux profils Ouest-Est d'une côte à l'autre et Nord-Sud une précision de ± 4 cms. pour une distance mesurée. Des résultats glaciologiques ne pourront être déduits de ces mesures qu'après leur répétition, prévue au plus tôt dans 4-5, au plus tard dans 8-10 ans. Déjà en 1959, le mouvement différentiel d'une section du profil Ouest-Est près de la côte Ouest fut déterminé par répétition des mesures. La section de 35 kms. a subi une extension totale de 9.10 ms. durant 3 mois. — Le Telluromètre a démontré être propre à la détermination des points fixés aux névés avec haute précision.

The plan to determine the position of fixed points in the Greenland Ice Cap by distance measurement with the Tellurometer originated from the experiences of former expeditions. The geodetic activities of both the German Greenland Expedition 1929-31 under Alfred Wegener and the French Greenland Campaigns of the Expéditions Polaires Françaises under Paul Emile Victor had proved that any measurement of angles on the Ice Cap meets with great difficulties due to the meteorological and climatic conditions, and provides only a limited accuracy.

The International Greenland Expedition (EGIG = Expédition Glaciologique Internationale au Groenland) had demanded for its investigations of the budget and the dynamic of the Ice Cap the determination of stick profiles with a relative accuracy of ca. 10 cm in position. This high accuracy could never have been achieved with the «classical» methods of triangulation or polygonisation in the short period of one summer campaign and with the limited means of EGIG. Therefore, a method was planned and prepared which avoided totally the measurements of angles and, at the same time, provided the running control of the measurements immediately on the field. The measurements of quadrangles with diagonals—as shown in Fig. 1— by Tellurometer was regarded as the most simple method of this kind. The Tellurometer had been proved in the preceding years as a reliable instrument. Its aptitude for

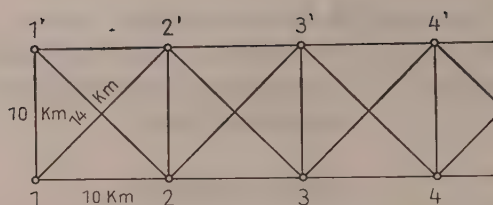


Fig. 1 — Chain of quadrangles with diagonals for tellurometer measurement.

measurements over snow-capped surfaces was known by applications in the arctic Canada. It was confirmed by experiments at Jungfraujoch, at Zugspitzplatt and in the surroundings of Munich, arranged by EGIG during autumn 1958 and winter 1958/59. The last mentioned experiments tested also the measurements of quadrangles with diagonals, giving for every quadrangle one surplus determination which can be used for immediate control.

For Greenland—and generally for any polar region—this method offers the following advantages:

1. The measurement becomes to a large extent independent of the surface relief, as the electromagnetic waves of the Tellurometer nestle in slight undulations of the terrain. They are disturbed or reflected only by steep slopes or walls which are not found on the Ice Cap.

2. High humidity of the air up to the formation of fog does not disturb the electric distance measurement. On the contrary, it raises the reliability of the distance determination out of the running-time of the waves by the greater homogeneity of the lower air layers in comparison with radiant weather conditions. Even slight snowfall does not prevent measurements. Thereby, the measurements become considerably independent of the weather. They are delayed only by strongly marked periods of bad weather or storm.

The distance measurement with Tellurometer and in the described arrangement proved soon to be an excellent method for the settlement of a net or profile of fixed points on the Ice Cap. The physical conditions at the Ice Cap are especially favourable for the Tellurometer measurements. Indeed, the electromagnetic waves suffer of a very high absorption when going close by the firn surface. Hereby, the range of the measurements is limited to about 10-12 km. On the other hand, just this strong absorption eliminates nearly completely the reflexion of the waves from the ground. Measurements with different frequencies cause but slight changes in the time readings and show nearly no swing which otherwise in size and course must be carefully considered when analysing Tellurometer measurements. In series with 10 frequencies, throughout carried out in Greenland, the swing reached in only a few cases more than one unit ($1 \mu\text{sec}$), a value which can be interpreted obviously with the insecurity of the readings and with real changes in the running-time.

To this high precision in the determination of the running-time intervenes the great reliability which can be attributed to the determination of the refraction index and with that of the actual velocity of light over the Ice Cap. The fall-wind, normally coming down from the summit of the Ice Cap to its borders, causes homogeneous meteorological conditions in the lower air layers. Therefore, the meteorological field along a certain distance can be described with sufficient accuracy by measurement of the temperature, pressure and humidity in the final points. With this, the insignificance of the vapor pressure is of special importance. In spite of a relative humidity which reaches mostly 80-90% the atmosphere contains due to its low temperature but

very little vapor whose pressure rarely transgresses 1-2 mm Hg. Only during calm and sunny days, over the surface of the Ice Cap a sharply limited inversion layer is developed. On such days, not only the meteorological observation in the lower layers were extremely insecure and questionable, rather the Tellurometer measurement itself gave very unreliable, or no values, as over unfavourable distances the measuring waves did not reach the other station—by reason of refraction or reflexion.

Under these conditions, the accuracy of the measured distances can be expected to correspond to the inner accuracy of the Tellurometer. The controls in the measured complete quadrangles show that this accuracy was reached in fact in Greenland. The differences between measured and computed distances were from the start surprisingly small. Their average for all complete quadrangles in the West-East-Profile was ± 12 cm. Hence follows an accuracy of ± 4 cm for a measured side.

Of course, the strong absorption of the measuring waves by the surface firn causes difficulties which prevented the planned establishment of a completely regular chain of squares, as sketched in Fig. 1. The surface of the Ice Cap is traversed by wavelike ridges and valleys,—especially in the marginal zones—whose course is very irregular. It proved to be impossible to surmount such ridges with the Tellurometer measurements. Rather, the stations had to be established on the ridges themselves whose distances changed between 4 and 10 km. Hence resulted in the marginal zones a chain of quadrangles whose links departed considerably from the intended form of squares. Furthermore, in spite of careful reconnaissance, it was not always possible to measure both diagonals of a quadrangle. In such cases, angles were measured for control.

In the real marginal zones of the West-East-Profile, namely between the Danish fixed point Qapiarfik and Carrefour in the West and between Depot 420 and Cecilia Nunatak in the East, and in the whole North-South-Profile, polygonal courses were measured ⁽¹⁾. The very undulated relief of the surface would have rendered there the establishment of a chain of quadrangles very difficult. Furthermore, crevassed regions forced the measurement to be confined to a single and well acquainted line. The measurement of the angles with the Theodolite Wild T 3 was throughout possible, but was delayed again and again by bad visibility or abnormal refraction. For control, the distances were measured twice independently with two Tellurometer equipments.

The North-South-Profile from Terme EGIG over Carrefour to T 132 (40 km north of Point S) crosses the upper regions of several big glaciers flowing down to the Fjords of the West Coast. It can be expected that the observations in this profile give special informations on the distribution of the glacial flow of these glaciers.

Altogether, during the summer campaign of EGIG the following distances were measured by Tellurometer and marked with sticks in both profiles West-East and North-South:

West-East-Profile Qapiarfik—Cecilia Nunatak:

Length of chain of quadrangles:	651 km
Length of polygonal courses:	274 km
Total Length:	925 km
Number of stations in the profile:	104
Number of sticks:	76
Sum of distances, measured by Tellurometer:	4100 km

North-South-Profile Terme EGIG—T 132:

Length of polygonal course:	280 km
Number of sticks:	35

⁽¹⁾ For all local citations see the map of the Expedition area in the article of Mälzer and Möller, page 474

Of course, glaciological results can be derived from the measurements in Greenland only after their repetition which is planned at the earliest in 4-5, at the latest in 8-10 years. This repetition will give full information both on the relative shifting between the sticks and on their absolute movement towards the ice edge, i.e. on the relative and absolute dynamic of the surface of the Ice Cap in the measured profiles. The fact that the Tellurometer measurements on principle is suitable to comprehend these movements with high accuracy was already proved by repeated measurements during the summer campaign 1959. The West-East-Profile between the points *T4* (Carefour) and *T301* (8 km west of Camp VI) was measured again, 3 months after the first measurement. It consists of 4 sections of approximately equal length; its overall distance is 35 km. The following table contains the distances of both the first and second measurement:

Point	Distance May 13, m	Sum 59 m	Distance August 13, m	Sum 59 m	Expansion m
<i>T301</i>		0		0	
	9425,16		9426,71		1,55
<i>T1</i>		9425,16		9426,71	
	7930,89		7936,27		5,38
<i>T2</i>		17356,05		17362,98	
	8669,41		8669,90		0,49
<i>T3</i>		26025,46		26032,88	
	9275,66		9277,34		1,68
<i>T4</i>		35301,12		35310,22	
				Sum:	9,10

The distance between *T301* and *T4* was affected in 3 months by a total expansion of 9,10 m which however is distributed very irregularly to the respective sections.

Also from the geodetic point of view, only the repetition of the measurements will give final results, as the sum of the distances from coast to coast is falsified by the movement of the ice during the 3 months of the measurements in the West-East-Profile. This movement is unknown until now and will be determined only by the repetition.

The experiences of EGIG have proved clearly the aptitude of the Tellurometer measurements for the establishment of glaciological profiles. The unexpected high accuracy achieved in Greenland makes it reasonable to repeat the measurements already after a short interval even when the movement of the ice is relatively small. The measuring technique and its performance with Weasels have been confirmed in principle. However, most of the working time on the Ice Cap was spent purely driving from station to station. Therefore one may expect a considerable speeding up of the measurements if instead of the slow Weasels a quicker vehicle, for instance helicopters, are combined with the Tellurometer. Preliminary considerations prove such a combination to be possible from the technical and operational point of view. Further work may confirm this opinion.

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DAS NIVELLEMENT BEI DER EXPEDITION GLACIOLOGIQUE INTERNATIONALE AU GROENLAND (EGIG) — SOMMER-KAMPAGNE 1959

Hermann MÄLZER und Dietrich MÖLLER (Karlsruhe)

SUMMARY

During the Expédition Glaciologique Internationale au Groenland (E. G. I. G.) in 1959 was performed for the first time a geometrical levelling in the arctic in order to obtain the vertical ice movements. This idea was given by Prof. Dr. H. LICHTE, Karlsruhe, who pointed out that the trigonometrical levelling, employed up to now, cannot give sufficiently exact results, because of the high refraction disturbances. The accuracy of the absolute altitudes of the bench marks (balises) on the inland ice should not exceed ± 1 m. The measurements were prepared by Dr. W. HOFMANN and performed by the geodetic group « Nivellement » under the collaboration of W. L. FROMMER, K. SCHNÄDELBACH, and the authors.

The levelling was joined with the point A 14, the altitude of which is 560,4 m above sea level of the western coast, and was carried out in the West-East-Profile of the EGIG by Disko Bugt via Camp VI EGIG, Milcent, and Station Centrale to the Station Jarl-Joset nearly 700 km. The work began on May 11, and the Station Jarl-Joset was reached on July 28. The elevation transmission was performed on foot across the 65 km broad crevasse area in the west of Groenland, and on the inland ice by means of vehicles (weasels). The measurements were performed by two independent parallel levellings and are therefore checked. The altitudes were determined along the profile of 65 bench marks. The walk speed amounted to an average of 7,2 km/day in the crevasse area, and on the inland ice 11,8 km/day. Automatic levelling instruments of the type « Zeiss Ni 2 arctic » and invar staffs were used for the observations. The geometrical levelling proved highly efficient. The mean error of a 1 km levelling is about $\pm 1,5$ cm. The differences of the altitudes between balises, approximately 12 km distant, are determined with an accuracy of 4-6 cm.

Besides the levelling there were executed barometrical levellings according to different methods in order to study the relevant range.

1. VORBEMERKUNG

Das zentrale Problem der EGIG ist die einheitlich ausgerichtete glaziologische Erforschung des Inlandeises als des größten Gletschers der Arktis, der mit 2,6 Millionen Kubikkilometer rund 12% des gesamten Eisvolumens der ganzen Erde umfaßt. Zur Betrachtung des Massenhaushaltes sowie zur Klärung der Zusammenhänge zwischen den grundsätzlichen Fragen der Mechanik, Thermodynamik von Schnee und Eis und der Verformungs- und Spannungsverhältnisse im Akkumulations- und Ablationsgebiet sind neben den glaziologischen, geophysikalischen und meteorologischen Beobachtungen geodätische Messungen erforderlich. Die geodätischen Arbeiten schaffen als Grundlage für die glaziologische-geophysikalische Forschung ein Punktfeld, das der Dynamik des Schnees und Eises ausgesetzt ist. Die zum Zeitpunkt der ersten Punktbestimmung nach Lage- und Höhe ermittelten Werte gestatten beim Vergleich mit den Ergebnissen der Wiederholungsmessungen Rückschlüsse auf horizontale und vertikale Bewegungen.

Das Operationsgebiet der EGIG liegt in einer Zone stärkster glazialer Aktivität und erstreckt sich im wesentlichen auf 2 Profile zwischen 69° und 73° nördlicher Breite. Das über 900 km lange West-Ost-Profil von der Disko Bugt über Station Centrale bis Cecilia Nunatak verläuft etwa über die gleiche Hauptachse, die von den Expéditions Polaires Françaises (EPF) 1948-1953 in vielen Einzelheiten untersucht ist. Ein zweites Profil verläuft in nord-südlicher Richtung etwa 100 km vom westlichen Randgebirge entfernt.



KARTE

Expédition Glaciologique Internationale au Groenland, Campagne d'été 1959.

- Anmarschweg vom Flughafen zum Operationsgebiet;
- Nivellement im Ost-West-Profil;
- Übriges Operationsgebiet ohne Nivellement.

2. AUFGABE

Aufgabe der Arbeitsgruppe Nivellement war es, auf dem West-Ost-Profil durch das im Ablationsgebiet westlich Camp Séismique bei der Vorexpedition 1958 von A. Baue gesetzten Ablationspegel und anschließend die im Akkumulationsgebiet des Inlandeises von der Gruppe Lagemessung vermarkten und bestimmten Festpunkte (Balisen) höhenmäßig einzumessen. Der Höhenanschluß sollte an den auf dem gewachsenen Fels in Höhe der Disko Bugt (Ata Sund) gelegenen Punkt A 14 erfolgen, dessen Höhe über dem Meerespegel (560,4 m) bereits 1948 trigonometrisch durch A. Baue bestimmt wurde [1]. Nach den Forderungen der Glaziologie und Geophysik sollten die Höhen über NN der einzumessenden Festpunkte nicht mehr als ± 1 m unsicher sein. Um die relativen vertikalen Eisbewegungen möglichst genau zu erfassen, dürften sich aber erst durch Vergleich mit späteren Wiederholungsmessungen ergeben werden, was für die Höhenunterschiede benachbarter Punkte eine größere Genauigkeit anzustreben. Die Wahl des Meßverfahrens war in erster Linie abhängig von den gestellten Genauigkeitsforderungen.

3. VORUNTERSUCHUNG UND WAHL DES MESSVERFAHRENS

Für die Höhenübertragung auf dem grönländischen Inlandeis wurden bisher das barometrische und das trigonometrische Meßverfahren eingesetzt. Obwohl die barometrische Höhenmessung bei Expeditionen stets eine bedeutende Stellung eingenommen hat und auch in Zukunft einnehmen wird, wenn es gilt, Höhen mit hinreichender Genauigkeit zu bestimmen, konnte sie in diesem Falle den hohen Anforderungen keinesfalls entsprechen. Den Anwendungsbereich der trigonometrischen Höhenmessung hat H. Lichte vor allem auf Grund der umfangreichen Messungen der EPF eingehend untersucht [2]. Die örtlichen Refraktionskoeffizienten im Meßbereich von 1 bis 2 m über Eis schwankten um mehr als das 200-fache der mittleren wirksamen Refraktionskoeffizienten $k = 0,13$.

Weiterhin ließen die Untersuchungen erkennen, daß die in Station Centrale ermittelten örtlichen Refraktionskoeffizienten aber auch bis zum 5-fachen des über dem Inlandeis wirksamen Refraktionskoeffizienten anwachsen. Diese Tatsache veranlaßten H. Lichte zur Durchrechnung verschiedener Zielstrahlen. Es zeigte sich, daß ein im Höhenbereich von 1,2 bis 2,0 m etwa parallel zur Schneeoberfläche verlaufender Zielstrahl tagsüber sinusartige Schwingungen mit Wellenlängen von einigen Kilometern und Amplituden von wenigen Dezimetern beschreibt, wodurch sich eine plausible Erklärung für die verschiedenen Größenordnungen des über eine größere Strecke wirksamen Refraktionskoeffizienten und des entsprechend der örtlichen Refraktionskoeffizienten ergibt. Wegen der nicht erfaßbaren Asymmetrie des Zielstrahlenverlaufes ist bei gleichzeitiger und gegenseitiger Beobachtung auf Zielweiten von nur 1 Kilometer mit einem mittleren Fehler des Einzhöhenunterschiedes von $\pm 0,04$ m zu rechnen. Unter Berücksichtigung der weiteren unregelmäßigen Messungsfehler (Zenitdistanz, Strecke, Instrumenten- und Zielhöhe) wäre bei beiderseitigem Anschluß an fehlerfreie Höhen für einen in 3000 m Höhe in der Mitte des Inlandeises gelegenen Punkt (etwa 450 km von den Anschlußpunkten entfernt) ein mittlerer Höhenfehler von ± 1 m zu erwarten. Mit durchschnittlichen Zielweiten von 1 km Länge, die relativ genau — etwa mit Basislatte — bestimmt werden müssen, wären theoretisch die gestellten Genauigkeitsanforderungen erfüllen. Die Verteilung der Höhenmessung auf mehrere Jahre ist wegen der nicht erfaßbaren vertikalen Eisbewegungen unmöglich; die trigonometrische Höhenüberbrückung des Inlandeises in einer Sommerkampagne scheitert jedoch an den hohen personellen und materiellen Aufwendungen.

Auf Grund dieser Überlegungen schlug H. Lichte vor [2], durch ein geometrisches Nivellement mit Zielweiten von etwa 100 m bei gleichzeitigen und gegenseitigen Beobachtungen mit automatisch horizontierenden Nivellierinstrumenten und Fahrzeugunterstützung die Genauigkeit der Höhenbestimmung zu steigern und die Arbeitsgeschwindigkeit zu beschleunigen. Diesen Vorschlag von Professor Dr. H. Lichte folgend, beschloß das Direktionskomitee der EGIG unter dem Präsidium von Professor Dr. R. Finsterwalder, das geometrische Nivellement im Rahmen der EGIG erstmalig unter arktischen Verhältnissen durchzuführen [3].

4. VORBEREITUNG

Die Vorbereitung, die volle 2 Jahre in Anspruch nahm, erstreckte sich auf die Erprobung des vorgeschlagenen Meßverfahrens und der zweckmäßigsten instrumentellen Ausrüstung. Unter dem Leiter der geodätischen Arbeitsgruppen (Lagemessung und Nivellement) Dr. W. Hofmann wurden Versuchsmessungen auf Alpengletschern^[4] und in Fontainebleau durchgeführt, wo die für die Expedition vorgesehenen Raupenfahrzeuge (Weasel) bereitgestellt wurden. Die Ergebnisse lieferten bei Zielweiten von 100 m einen mittleren zufälligen Kilometerfehler für das Doppelnivellement bis zu ± 5 mm. Es war aber zu erwarten, daß diese relativ hohe Genauigkeit unter den wesentlich härteren Bedingungen in Grönland nicht einzuhalten ist. Die Zeitstudien ließen erkennen, daß in den wenigen zur Verfügung stehenden Sommermonaten (Anfang Mai bis Mitte August) die einmalige und vollständige nivellitische Überbrückung des 900 km langen West-Ost-Profiles für eine Arbeitsgruppe von 4 Geodäten nur unter äußerst günstigen Umständen durchführbar ist. Auf alle Fälle bestand die berechtigte Aussicht, im Hauptarbeitsgebiet der Glaziologie und Geophysik von der Westküste über Camp VI EGIG und Station Centrale bis zur Überwinterungsstation Jarl-Joset ein Höhenprofil mit einer Genauigkeit zu gewinnen, die durch kein anderes Meßverfahren erreichbar ist.

Die instrumentelle Ausrüstung wurde in dankenswerter Weise vom Deutschen Geodätischen Forschungsinstitut München und Frankfurt, vom Institut für Photogrammetrie der Technischen Hochschule München, vom Geodätischen Institut der Technischen Hochschule Karlsruhe und von der Deutschen Forschungsgemeinschaft zur Verfügung gestellt.

5. FUSSNIVELLEMENT

Nach Aufbruch zur Sommerkampagne Anfang April 1959 und nach Errichtung des Lagers Camp VI EGIG begannen die wissenschaftlichen Arbeiten am 11. Mai. Als Wissenschaftler gehörten der Arbeitsgruppe Nivellement W.L. Pfrommer (Karlsruhe), K. Schnädelbach (Bergzabern/Pfalz) und die Verfasser an. Auf Grund der Erfahrungen der EPF mußte die Höhenübertragung im Ablationsgebiet durch die Bruch- und Spaltenzonen unter Verzicht auf die Raupenfahrzeuge zu Fuß durchgeführt werden. Die von A. Bauer vor Beginn der Messung durchgeführte Erkundung und Absteckung der Marschroute schufen die Voraussetzungen für das Gelingen des Fußnivellements und ergaben, daß der Zusammenschluß des Nivellements durch das Randgebiet und des eigentlichen Inlandeisenivellements nicht wie ursprünglich vorgesehen bei Camp VI EGIG notwendig war, sondern an dem etwa 30 km westlich gelegenen Punkt Camp Séismique erfolgen konnte. Zur Unterstützung der 4 Geodäten haben als Lattenträger der Arzt der Überwinterungsstation H.G. de Sypiorski und O. Schimpp von der Gruppe Küstenglaziologie am Fußnivellement teilgenommen. Während des Fußmarsches war die Gruppe Nivellement völlig

auf sich angewiesen und mußte neben der instrumentellen und lebensnotwendigen Ausrüstung 2 Zelte auf Handschlitten mitführen. Wegen der Spaltengefahr mußte der größte Teil der Arbeiten unter Seilsicherung durchgeführt werden.

Das Nivellement wurde als Vierfachnivellement, das einem unabhängigen Doppelnivellement entspricht, mit 2 Ni 2 arctic auf starren Stativen und 2 4m-Klappplatten (E-Teilung) durchgeführt. Die Messungsanordnung ist in Abbildung 1

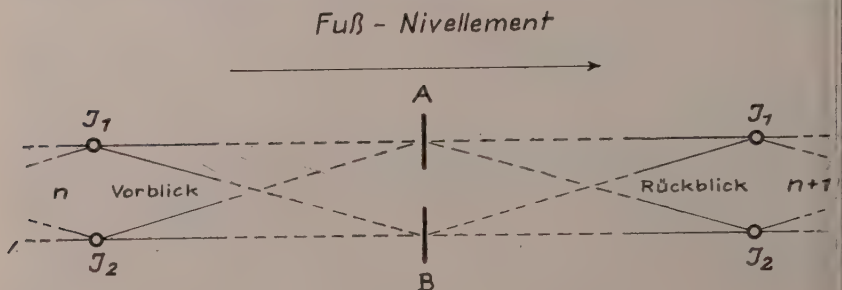


Abb. 1

schematisch dargestellt. Mit den beiden Instrumenten I_1 und I_2 wurden die beiden getrennt stehenden Latten A und B abgelesen, wobei sich als Standpunktkontrollen für den Vorblick $\Delta V_t = V_A - V_B = \Delta V_2$ und für den Rückblick $\Delta R_1 = R_A - R_B = \Delta R_2$ ergaben. Außerdem konnte die Standfestigkeit der Latten während des Instrumentenwechsels vom Standpunkt n zum Standpunkt $n+1$ dadurch kontrolliert werden, daß $\Delta V_n = \Delta R_{n+1}$ sein mußte. Um ein befürchtetes Einsinken der Instrumente während des Lattenwechsels zu erfassen, wurde ein in etwa 10 m von den Instrumenten in den Firn eingerammter Kontrollstab vor Beginn und nach Beendigung der Beobachtungen abgelesen. Es zeigte sich, daß weder die Latten, die auf einem Knopf zweier Holzplatten von 40 cm Durchmesser aufgesetzt wurden, noch die Instrumente, deren Stativfüße mit Schneetellern von 20 cm Durchmesser ausgerüstet waren, während einer maximalen Standzeit von 15 Minuten ihre Lage änderten.

Die ursprünglich vorgesehenen Zielweiten von 100 m konnten nicht eingehalten werden, da die Schnee- und Eisoberfläche ausgeprägte Täler und Höhenrücken mit Geländeneigungen bis zu 10% auswies. Je nach den Geländebedingungen ergaben sich Zielweiten von 15 bis 120 m. Um instrumentelle Fehler auszuschalten, wurden auf gleiche Zielweiten im Vor- und Rückblick besonders geachtet. Die Aufschreibungen der Beobachtungen erfolgte für jedes Instrument getrennt. Beobachter und Schreiber wechselten ihre Tätigkeit alle 2 Tage.

Bei Abbruch der Messung auf freier Strecke zwischen den Ablationspegeln, zum Beispiel bei Windgeschwindigkeiten über 10 m/sec oder bei starker Schneetrift, wurden durch Eingraben der Holzplatten und tiefes Einrammen des Kontrollstabes in den Firn Zwischenfestpunkte geschaffen, deren relative Höhenunterschiede in einem Zeitraum bis zu 2 Tagen unverändert blieben. Die am 16. Mai plötzlich eingetretene Schneeschmelze zwang die Gruppe zur möglichst schnellen Durchquerung der sich bildenden Seen und Gletschersümpfe. Um die Messungen mit der notwendigen Sicherheit fortzuführen, wurde ein großer Teil der noch zurückzulegenden Strecke bei Mitternachtsonne gemessen, da nachts das Schnee- und Eisfeld besser gangbar war. Die letzten Kilometer durch das Randgebiet wurden in 22-stündiger ununterbrochener Arbeit zurückgelegt.

Nach 13 Tagen, am 23. Mai, und nach Überwindung eines Höhenunterschiedes von 788 m konnte das Fußnivellement an dem durch 2 Höhenbolzen versicherten

Festpunkt A 14 auf dem festen Fels erfolgreich abgeschlossen werden. Unter Abzug von 4 Schlechtwettertagen wurde die 65 km lange Strecke mit einer durchschnittlichen Leistung von 7,2 km pro Tag zurückgelegt. Das Ergebnis wird die absoluten Höhen von 10 Ablationspegeln liefern.

6. INLANDEISNIVELLEMENT

Für die motorisierte Durchführung des Inlandeisnivellements waren der Gruppe 1 Mechaniker und 1 Funker als Fahrer zugeteilt. Die Arbeiten begannen am 28. Mai mit dem Höhenanschluß an die bei Camp Séismique zu Beginn des Fußnivellements gesetzte und versicherte Balise. Die Höhenübertragung wurde als unabhängiges Doppelnivellement (Vierfachnivellement) mit 2 Ni 2 arctic und 2 durch Kugelgelenkhalterungen an den Weaseln angebrachten 3m-Invarlatten mit Schachbrettteilung (siehe Abbildung 2) durchgeführt. Die Fortbewegung erfolgte mit Weaseln im

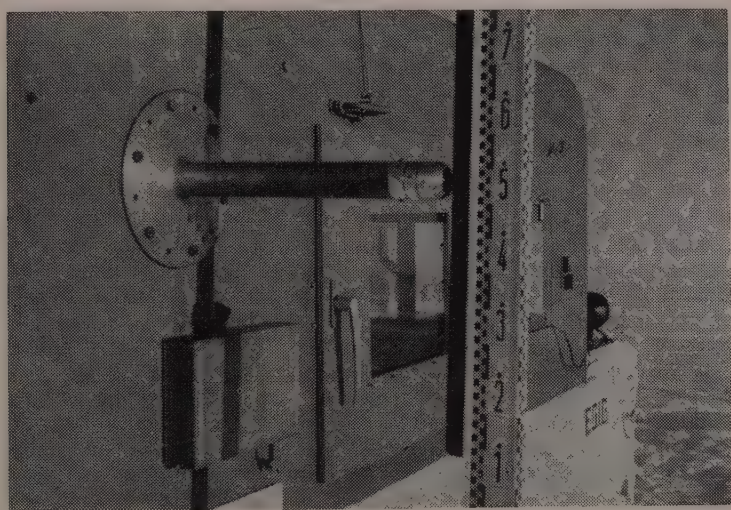


Abb. 2

Weasel - Nivellement

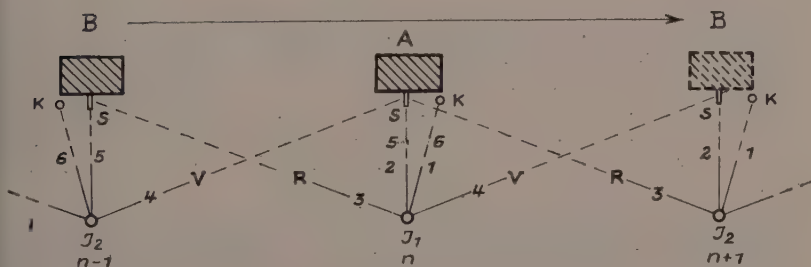


Abb. 3

überschlagenden Einsatz. Die Messungsanordnung und die Reihenfolge der Beobachtungen werden in Abbildung 3 wiedergegeben. Die beiden Beobachter stellten die Instrumente senkrecht zur Fahrtrichtung in einem Abstand von 10 bis 20 m gegenüber den Fahrzeugen auf. Der Beobachter A am vorderen Weasel begann mit der Ablesung des Kontrollstabes $1K_n$, dann folgte die Ablesung an der Stirnteilung der eigenen Weasellatte $2S_n$ und schließlich wurde über die Fernsicht der Rückblick $3R_n$ genommen. Zur gleichen Zeit beobachtete der Beobachter B beim hinteren Fahrzeug den Vorblick $4V_{n-1}$, die Stirnteilung der eigenen Weasellatte $5S_{n-1}$ und den Kontrollstab $6K_{n-1}$. Durch die gegenseitige Beobachtung beider Weasellatten ergab sich eine Standpunktkontrolle $\Delta H_{n-1} = (V - S)_{n-1} = (S - R)_n = \Delta H_n$. Die Höhenunterschiede ΔH wurden von den Schreibern, die in den Fahrzeugen saßen, durch Funksprechgeräte (AN/PRC 6) ausgetauscht und überprüft. Die Vergleichsunterschiede schwankten bei justierten Instrumenten zwischen 0 und 3 mm. Erst nach vorgenommener Kontrolle fuhr das jeweils rückwärtige Weasel vor, wobei es das vordere Weasel um die gleiche Entfernung «übersprang» (gleiche Zielweiten). Die Entfernungen wurden mit einem eingebauten Schrittzähler gemessen (62 Einheiten = 100 m).

Die Kontrollablesungen waren erforderlich, um ein Einsinken des Instrumentes und der Latte während des Sprunges zu erfassen. Es hat sich gezeigt, daß die Instrumente während des durchschnittlich 9 Minuten dauernden Standes ihren Horizont beibehielten, dagegen die Fahrzeuge und damit die Latten innerhalb weniger Minuten bis zu 3 mm einsanken. Die Beobachter saßen während der Fahrt mit dem Instrument in der offenen Weaseltür oder auf dem mitgeführten Schlitten bzw. in der offenen Tür des Wohnschlittens. Schreiber und Beobachter wechselten sich in ihrer Tätigkeit alle 2 Tage ab. Die Übermittlung der Meßwerte vom Beobachter zum Schreiber konnte auch bei extremen Witterungsverhältnissen (-30°C und starkes Schneefegen) durch Zuruf erfolgen, da die Weaseltüren ständig offen gehalten wurden und die Fahrzeugmotoren abgestellt waren. Die Beobachtungswerte wurden mit je 2 Handadditionsmaschinen (für Kontrollberechnungen und laufende Höhenberechnungen) vorgenommen. Die Zielweiten beim Nivellement waren sehr verschieden und richteten sich nach den Geländeverhältnissen. Vom Camp Séismique bis Milcent zeigte die Schneeoberfläche großräumige Wellenformen mit Steigungen und Gefälle bis zu 5% und noch darüber. Selbst bei Ausnutzung der gesamten Lattenlängen konnten öfters nur Zielweiten von 25 m genommen werden. Im Inneren des Inlandeises war nur noch eine flache Steigung von 0,2 bis 0,5% bis zum höchsten Profilpunkt Dépôt 275 (3180 m über NN) vorhanden. In derselben Größenordnung lag auch das Gefälle bis zur Überwinterungsstation Jarl-Joset, die rund 200 km von der Ostküste entfernt liegt. Als günstigste Zielweiten bei flachem Anstieg und Gefälle erwiesen sich Entfernungen von 70 Zählereinheiten = 113 m. Diese Distanz garantierte bei relativ rascher Marschgeschwindigkeit noch eine gute Schätzung der Millimeter am Lattenbild.

Als Beitrag zu Refraktionsstudien wurden in Verbindung mit den 3 täglichen meteorologischen Terminen 3-Fadenablesungen während des gesamten Nivellements und außerdem in den Abend- und Morgenstunden durch besondere Messungsanordnungen zusätzliche Beobachtungen vorgenommen. Dabei zeigte sich, daß vorwiegend in den Abendstunden bei starker Sonneneinstrahlung, Windstille und positiven Temperaturgradienten eine Neigung des Zielstrahles von mehreren Millimetern auf 100 m Entfernung auftrat. Dieser störende Einfluß wird aber weitgehend durch gegenseitige und gleichzeitige Beobachtung aufgehoben.

Am 28. Juli erreichte die Gruppe Nivellement nach 62 Reisetagen über das Inlandeis und einer zurückgelegten Entfernung von 605 km ab Camp Séismique über Milcent (17. Juni) und Station Centrale (1. Juli) die Überwinterungsstation Jarl-Joset. Diese Entfernung wurde unter Abzug von 9 Schlechtwettertagen und

2 Aufenthaltstagen in Station Centrale mit einer durchschnittlichen Tagesleistung von 11,8 km/Tag zurückgelegt. Die maximale Arbeitsleistung betrug 16,6 km/Tag. Es wurden 55 Balisen höhenmäßig eingemessen. Der Höhenunterschied von Camp Séismique bis zum Scheitelpunkt betrug rund + 1830 m und vom Scheitelpunkt bis zur Station Jarl-Joset ca. — 310 m.

Nach dem Zeitplan erreichte die Arbeitsgruppe Nivellement mit 18 Tagen Verspätung die Station Jarl-Joset. Einen Zeitverlust von 10 Tagen brachten bereits die besonders ungünstigen Verhältnisse (chaotisches Gelände und schlechtes Wetter) beim Aufstieg der Transportgruppen zum Inlandeis und die Errichtung des Lagers Camp VI EGIG. Die großen Schwierigkeiten beim Fußnivellement und die Installation und Bereitstellung der Weasel für den Marsch durch das Inlandeis ergaben eine weitere Verzögerung von 8 Tagen. Das Inlandeisnivellement konnte ohne Zeitverlust planmäßig durchgeführt werden.

Die fast 3-wöchige Verspätung und die von den Transportgruppen und von der Gruppe Lagemassung erkundeten schwierigen Geländebeziehungen im Profilstück zwischen Dépôt 420 und Cecilia Nunatak führten zu dem Entschluß, das Nivellement in Station Jarl-Joset zu beenden. Der nahezu 150 km lange Abstieg von dem etwa 100 km nordöstlich Station Jarl-Joset gelegenen Dépôt 420 bis zum Cecilia Nunatak dürfte für das Nivellement wegen der terrassen- und wellenförmigen Absätze fast noch größere Geländeschwierigkeiten bieten, als es im westlichen Ablationsgebiet der Fall war. Unter diesen Verhältnissen wäre im weiteren Verlauf der Route bis Cecilia Nunatak wieder ein Fußnivellement erforderlich gewesen, und die Gruppe Nivellement hätte vor Mitte September das Ziel kaum erreicht. Eine Rückführung der Gruppe zu diesem späten Zeitpunkt über das Inlandeis bis zur Westküste erschien unmöglich, da sich schon im August die Wetterlage im allgemeinen merklich verschlechtert und der grönländische Winter einbricht. Eine Abholung der Gruppe an der Ostküste war im Operationsplan nicht vorgesehen und hätte auch kurzfristig nicht organisiert werden können. Ein direkter, von dem abgesteckten Profil der EGIG abweichender Durchgang in östlicher Richtung von Station Jarl-Joset zum nächstgelegenen Nunatak (etwa 150 km) war durch breite und gefährvolle Gletscherspalten versperrt. Eine Weiterführung des Nivellements von Station Centrale in nordöstlicher Richtung bis zum Dépôt 420 wäre zeitlich noch möglich gewesen, es wurde aber davon zugunsten barometrischer Höhenmessungen während des Rückmarsches über das Inlandeis Abstand genommen.

Obwohl die Höhenübertragung ihren Abschluß nicht an der Ostküste fand, an der ohnehin bisher noch kein Höhenfestpunkt mit Anschluß an den Meerespiegel Ostgrönland vorhanden ist, wurde durch dieses erste Nivellement, das in sich durch ein unabhängiges Doppelnivellement kontrolliert ist, für die glaziologischen Untersuchungen bereits eine wertvolle Grundlage geschaffen. Zu einem glaziologisch bedeutungsvollen Bestandteil wird das Nivellement aber erst dann, wenn in einigen Jahren Wiederholungsmessungen stattgefunden haben werden. Neben den relativen Höhenänderungen zwischen den Balisen, die zur Erfassung der Gletscherdynamik sehr wichtig sind, wird die absolute Höhenlage des Inlandeises zum Zeitpunkt des Nivellements von größter Bedeutung sein. Der Vergleich mit den absoluten Höhen eines späteren Nivellements wird zeigen, ob sich die Eisfläche in Grönland hebt, senkt oder gleicht.

4. GENAUIGKEIT

Erst nach den Wiederholungsmessungen, die die vertikalen Eisbewegungen klar erkennen lassen werden, wird eine exakte Untersuchung der absoluten Höhen- genauigkeit möglich sein. Man kann jedoch aus den vorläufigen Differenzen des unab-

hängigen Doppelnivellements ein Maß für die zu erwartende Größenordnung der inneren Genauigkeit des Nivellements erhalten. Diese Genauigkeit wird beeinflusst durch ein nicht mehr erfassbares Einsinken von Latte und Instrument, instrumenteller Fehler, Lattenteilungsfehler, Refraktionsstörungen und durch Ablesefehler, die infolge von Schätzungsfehlern und verschiedenen Beleuchtungseinflüssen hervorgerufen werden. Der hieraus resultierende und zu erwartende mittlere Kilometerfehler für das Doppelnivellement wird für das Fußnivellement durch das Spaltengebiet und für das Weaselnivellement über das Inlandeis in der Größenordnung von $\pm 1,5$ cm liegen. Somit dürfte der relative Höhenunterschied von Balise zu Balise, die 8 bis 17 km auseinanderliegen, auf 4 bis 6 cm genau bestimmt sein.

Dieser kurze Hinweis läßt schon erkennen, daß das geometrische Nivellement vor allem auf dem flach ansteigenden Inlandeis seine Berechtigung hat. Selbst in den steileren Randgebieten, in den Bruch- und Spaltenzonen, kann es erfolgreich und mit Genauigkeitsgewinn — wie die Durchführung beweist — eingesetzt werden, wenn auch die Messung stets mit technischen Schwierigkeiten und erheblichen körperlichen Anstrengungen verbunden sein wird. Nur in Ausnahmefällen, wenn es gilt, breite Spalten zu überqueren oder sehr steil abfallende Bruchstreifen zu überwinden, wird man die trigonometrische Messung als Hilfsmessung anwenden, aber dann auch nur mit möglichst kurzen Zielweiten.

8. BAROMETRISCHE HÖHENMESSUNGEN

Um den Anwendungsbereich und die verschiedenen Verfahren der Barometrie im Polargebiet näher zu untersuchen, haben die Arbeitsgruppen Geophysik, Lagemessung und Nivellement im Rahmen ihrer Arbeitsprogramme zusätzlich barometrische Höhenmessungen durchgeführt. Für diese Untersuchungen liefern aus dem West-Ost-Profil die durch das geometrische Nivellement bestimmten Höhen exakte Vergleichswerte.

Der Arbeitsgruppe Nivellement standen zur Durchführung dieser Arbeiten 6 Aneroide, davon 5 Bodenhöhenmesser von Thommen und 1 Aneroid von Fuesse zur Verfügung. Das mit Verspätung angetretene Inlandeisnivellement dürfte durch die zusätzlichen barometrischen Messungen keine weiteren Verzögerungen erfahren. Da ein verlängerter Aufenthalt von nur 1 Minute pro Standpunkt (etwa 6000 Standpunkte von Camp Séismique bis Station Jarl-Joset) einen Zeitverlust von 10 Arbeitstagen ergeben hätte, konnte nur auf jedem zehnten Nivellementsstandpunkt die Ablesung aller Aneroide erfolgen. Die Lufttemperaturen wurden dabei mit 2 Schleuderthermometern gemessen und der absolute Luftdruck einmal täglich mit einem Hypsometer bestimmt. Das barometrische Nivellement wurde an die 4 Stationsbarographen in den Profilhauptpunkten Camp VI EGIG, Milcent, Station Centrale und Station Jarl-Joset angeschlossen. Für die Auswertung stehen zusätzlich Registrierungen der dänischen meteorologischen Küstenstationen zur Verfügung. Das Ergebnis wird zeigen, mit welcher Genauigkeit durch einfachste Messung die Höhenfestlegung von Marschrouten in Arktis und Antarktis erfolgen kann.

Auf dem Rückweg über das Inlandeis, der am 3. August begann, wurde auf der 262 km Strecke von Station Jarl-Joset bis Station Centrale das barometrische Sprungverfahren angewendet. Die Beobachtungsstandpunkte lagen in Abständen von 4 bis 5 km. Die Temperaturen wurden mit 2 Aspirationspsychrometern gemessen. Dieses Verfahren dürfte durch die Erfassung der zwischenzeitlichen Luftdruckschwankungen mit Standbarometern und durch die gleichzeitige Ablesung auf benachbarten Standpunkten sowie durch den laufenden Vergleich der Aneroide untereinander die genauesten barometrischen Ergebnisse erwarten lassen. Bei einer durchschnittlichen Marschleistung von 35 km/Tag erreichte die Gruppe am 10. August Station Centrale.

Wegen des kurzbevorstehenden Abtransportes vom Inlandeis mußte die Marschgeschwindigkeit gesteigert werden und deshalb wurde die Messung nach der Parallelmethode fortgesetzt, wobei sich ein durchschnittlicher Arbeitsfortschritt von 66 km/Tag ergab. Am 15. August mußten nach Behebung schwerer Fahrzeugschäden die Messungen 70 km vor Camp VI EGIG abgebrochen werden, um das Lager für den Abtransport noch termingerecht zu erreichen.

9. ZUSAMMENFASSUNG

Bei der EGIG wurde zum ersten Male das geometrische Nivellement für eine genaue Höhenbestimmung mit Erfolg eingesetzt. Obwohl für die Vorbereitung und praktische Durchführung dieser Messungen bisher noch keine Erfahrungen vorlagen, konnten in rund 100 Tagen fast 700 km nivellitisch überbrückt werden. Nach den vorläufigen Berechnungen übertrifft die innere Genauigkeit des Verfahrens die Forderungen der Glaziologie und Geophysik.

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THE USE OF AERIAL PHOTOGRAMMETRY IN THE STUDY OF MOUNTAIN GLACIERS

T. J. BLACHUT

Photogrammetric Research
National Research Council
Ottawa : Canada

SUMMARY

A growing demand for the use of aerial photogrammetry in the study of glaciers prompted the author to discuss in some detail this new application of aerial photogrammetric technique. Based on experience gained from the Salmon Glacier research project and from the initial stage of the Axel Heiberg project, various phases of the application of aerial photogrammetry to the study of mountain glaciers are systematically reviewed. Emphasis is placed on the proper planning and execution of photographic flights and the plotting of glaciological maps, which are special-purpose maps governed by different rules from those followed in the production of the usual topographical maps.

RÉSUMÉ

La photogrammétrie aérienne est de plus en plus employée dans l'étude des glaciers de montagne. Cela a incité l'auteur à décrire de façon assez détaillée ce qui est une nouvelle application de la technique photogramétrique. Les divers aspects de cette application sont passés en revue, compte tenu des données acquises au Salmon Glacier (où l'on a expérimenté la nouvelle technique) et à Axel Heiberg (où l'on a commencé une étude systématique). L'auteur souligne l'importance de la préparation et de l'exécution des vols photographiques, et de l'établissement des cartes glaciologiques. Ce sont des cartes spéciales qu'on établit selon des règles différentes de celles auxquelles on a recours pour l'établissement des cartes topographiques ordinaires.

1. INTRODUCTION

In spite of the intensive development of aerial photogrammetry during recent years, its use in the study of mountain glaciers has been very limited. One reason for this limitation is the fact that until recently glacier studies have not been appreciated or sufficiently supported by state authorities. As a result, scientists in glacier research did not have access to the more efficient but also more expensive tools in their work. Today the situation has changed, and in some countries glacier studies have become part of the national scientific effort. With the financial support of government authorities it is likely that aerial photogrammetry will be more intensively used in glacier research.

On the expedition to Salmon Glacier in British Columbia in the summer of 1955 an attempt was made to use aerial photogrammetry in a more systematic way than had been done so far. The results were satisfactory. This year another project requiring intensive use of aerial photogrammetry is underway at the Hugh Thompson Glacier on Axel Heiberg Island in the Canadian Arctic. From the experience gained from both the above-mentioned projects, the author intends to discuss some basic aspects of the use of aerial photogrammetry in glacier studies.

2. THE USE OF AERIAL PHOTOGRAPHS IN PREPARING FOR GLACIER EXPEDITIONS

In planning glacier expeditions, it is most important to make use of existing aerial photographs, particularly if the glacier is in an unmapped area. The scale of photographs is a secondary matter, but for planning purposes the most recent photographs are preferred. From these photographs a *provisional map* of the expedition area should be produced. It is suggested that the plotting be performed at a scale of 1:25,000 with the *form line interval* ranging from 20 to 100 m, depending upon elevation differences and the ruggedness of the terrain. Because no ground control points may be available in unmapped territories, the mapping scale and the elevations will be rough approximations, but this will not seriously affect the value of the provisional map. It is sufficient to have the map in only one color, and a simple technique, such as "Ozalid", can be used to make copies.

For planning the ground control network it is very important to outline the limit of snow-covered areas very precisely on the provisional map.

The scale of 1:25,000 is considered to be most suitable for a map that is to be used for the field plotting of detailed data by geologists, geophysicists, botanists, and other members of the expedition. For the planning of field survey and photographic coverage and for the execution of photographic flights, a scale of 1:50,000 will be found more convenient. Provisional maps at this scale can be obtained by a simple photographic reduction from the 1:25,000 map.

In addition to provisional maps, a photomosaic of the expedition area should be produced. An original photomosaic can be produced in any scale, but reproductions in the approximate scale of 1:50,000 (1:100,000 for larger areas) would be most practical.

Provisional maps, accompanied by photomosaics and aerial photopairs to be viewed under a stereoscope, will permit very detailed planning of the field survey and photographic flights. A basic ground control net, together with a tentative observation program, can be designed prior to the departure to the field. Similarly, preparations for other expedition activities can be worked out in the office and the members of the expedition can familiarize themselves with the expedition area even before they reach the field. This careful planning and study of the expedition area from existing aerial photographs will save much valuable time on the actual mission and may help to avoid grave misconceptions about the general organization and conduct of expedition activities.

The production of a provisional map does not take much time or money and, therefore, should not be neglected if photographic coverage of the expedition area is available. For larger expeditions, however, *it should become a routine procedure* even if aerial photographs should have to be especially ordered for this purpose. Our experience gained so far in preparing for the Axel Heiberg photogrammetric mission indicated that even photographs of 1:60,000 are very suitable for preparatory work. In our case the plotting of a provisional map at scale 1:25,000 of the expedition area (25 km \times 40 km) required approximately two weeks' time for one operator.

If aerial photographs from earlier photographic missions are also available the should be studied and compared with the more recent photographs, since this comparison may show up areas which should be submitted to special investigation.

The use of aerial photographs in planning glacier expeditions offers obvious advantages, but it should not be limited to a simple examination of photo prints. With the photogrammetric facilities presently available in practically all countries, the plotting of a provisional map of the expedition area must be considered as an indispensable first step in glacier research projects.

3. THE PLANNING OF PHOTOGRAPHIC FLIGHTS

As far as the aerial photogrammetric technique is concerned, only very small areas are involved in mountain glacier studies. Therefore, a generous planning of the photographic coverage is permissible and indicated. Additional photo flights will add little to the cost, but they may be very helpful in the final measurements and mapping of the glacier and its environs.

General Coverage

First of all a general photographic coverage of the glacier and surrounding area should be considered. The scale of photographs depends upon the terrain elevation and the ceiling of the aircraft used, but for general coverage scales 1:40,000 and smaller are most convenient. On first-order plotters, and with modern 1:50,000 photography, an accuracy (mean square error) of ± 1.0 m is possible in the x - and y -coordinates and a slightly better accuracy in z . The horizontal accuracy quoted above is satisfactory for maps at 1:10,000 and possibly even for maps at 1:5,000. Therefore, the general photographic coverage at the suggested scales can be used for plotting a small-scale map of the general area as well as for establishing the secondary horizontal control that is required in the large-scale detail surveying and mapping of glaciers.

In high mountains an 80% longitudinal overlap is recommended.

Aerial Triangulation Strips

The determination of the necessary number of control points for each individual overlap of low-altitude photographs used in the measuring and mapping of glaciers may be difficult and also impractical, depending upon local conditions and the facilities available. Basic low-altitude photographs of glaciers may show irregular patterns and may not cover sufficient bare terrain to include primary ground control. Therefore, the flying of special photographic strips for aerial triangulation should be considered. In the main these strips will be flown along the axis of the larger mountain glaciers, but they may also be planned for other purposes as, for instance, the bridging of ice caps and snow fields on which no reliable fixed control points can be established. In the bridging of snow-covered areas, it is essential to provide a sufficient number of points which will be clearly visible on aerial photographs and which could be used as carry-over points in aerial triangulation. Black targets, or sawdust mixed with lamp-black, could be used to mark these points.

The scale of photographs and length of triangulation strips depends on the required accuracy and specific terrain conditions. In view of the stringent accuracy requirements in glacier work, only two to three overlaps can be bridged and this means that for every two to three overlaps, it is necessary to provide vertical control points at least for adjustment purposes.

Photographic Coverage for Detailed Measuring and Mapping of Glaciers

The most important photographs in glacier studies are those from which the actual measurements and the mapping of glaciers are to be carried out. Several factors must be carefully considered. One of them is of course the scale of the photographs, and in this regard two cases must be distinguished.

a) where glacier measurements must be referred to fixed points on solid ground, and

b) where the reference to fixed points on solid ground is not required or does not present any difficulty because of the limited size of the glacier.

a) If the measurements on a glacier must be tied into fixed points on solid ground, difficulties will arise from the fact that single photographs from low altitude may not

cover the entire width of the glacier with bands of bare ground on either side. The only other alternative is to establish control points on the glacier and to tie them into the solid ground points by means of field surveying. However, since the glacier is in constant flow, the fixing of control points on the glacier must be done during the flight operations or within a short time interval prior to or after the photo flights so that the glacier motion will not adversely affect the required accuracy of the control net. The signalization of points on the glacier surface may present additional difficulties.

Therefore, if it is at all possible, it is practical to make aerial photographs for basic glacier measurements from an altitude that will permit the imaging, on a single photograph, of the entire width of the glacier and its bordering bands of solid ground. This requirement may restrict the lower limit of flying height. Using modern photographic and plotting equipment, however, a remarkable accuracy of the order of 0.1‰ of the flying height can be obtained in the photogrammetric determination of spot elevations. This means that for 1:25,000 photographs taken with a modern camera of 153 mm focal length and 230 × 230 mm size, an accuracy of ± 37 cm is possible and the ground coverage would amount to 5.7 × 5.7 km, which will even permit coverage of the larger glaciers.

b) If there are no requirements for the tying-in of glacier measurements to control points on solid ground, or if this can be easily provided because of the small size of the glacier or by means of control points established on the glacier, photography from a very low altitude may be used.

Another factor which must be considered is the slope of the glacier and the configuration of the surrounding terrain. If the slope is very steep, it may be necessary to fly at several steps in order to maintain the desired relative flying height. Occasionally the lower limit of flying height may also be restricted by the manoeuvrability of the aircraft because of the limitations imposed by the landforms surrounding the glacier.

4. FIELD PREPARATION

Field preparation includes:

a) the establishment of a ground control net,

b) signalization for the purpose of photogrammetric evaluation of photographs, and

c) field identification.

a) The establishment of a ground control net does not require special explanation since the rules are the same as those for other photogrammetric work. Special attention must be paid to the control points located on glaciers since they change their position in all three coordinates with the movement of the glacier. Therefore, they must be established almost simultaneously with the photographic flight. Often only the ground elevations of a number of points are required, and their determination can be accelerated by measuring only the vertical angles in the field and using the *photogrammetric horizontal distances* in computations. From the slope of the sight lines involved, it is easy to decide whether or not this procedure is permissible in view of unavoidable small errors in the photogrammetrically-established distances.

b) The signalization of points located on solid ground does not present special difficulties. It should not be overlooked, however, that rock surfaces are highly reflective and show up as light gray or white spots on aerial photographs. As a result, there is very little contrast between a white or yellow target and bare rock, and the targets may not be recognizable on photographs. Therefore the contrast between the target and the background should be increased by, for instance, coating the surface around the target with a dark, unreflecting substance. At the same time, since a greater contrast improves image resolution much smaller targets will suffice.

Once sufficient contrast is secured, the size of targets for high resolution lenses can be determined from the diagram in Fig. 1. The establishment of signalized points in a shadow area should be avoided.

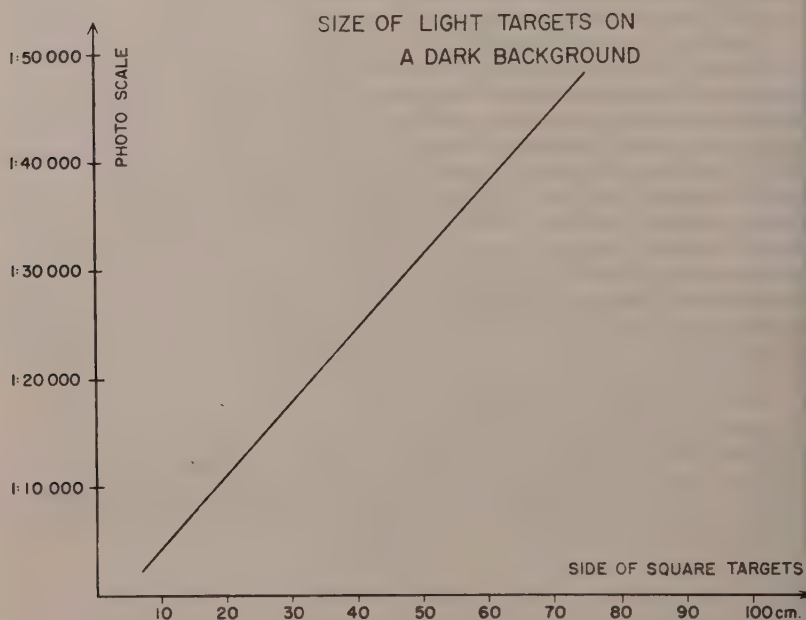


Fig. 1

There is another characteristic feature of rocky terrain which should be taken into consideration. Photographs of rocky ground often show many white spots which are very similar to target images. Therefore there is a constant danger of wrong identification, particularly since aerial photographs suitable for field identification work are seldom available in glacier work. It is therefore desirable to make the targets in a very distinctive form, for instance, in a cross.

The problem of signalizing points located in snow-covered areas and on the glacier is much more complex and only little experience has been obtained on this subject. The main difficulties result from the fact that the position of targets may be changed by the melting of snow or ice and that their visibility is limited by the excessive brightness of the surrounding area. At Salmon Glacier, for instance, not one of 30 black targets 60 cm \times 60 cm placed on the glacier was visible on aerial photographs made at a scale as large as 1:6000. Better results could probably be obtained with newer lenses, but in any case, the size of black targets on a light background must be at least three times that of a white target on a dark background.

Further investigations into the signalizing problem in glacier studies have been planned for this year's project at Axel Heiberg Island. In addition to experiments with flat black targets, it is intended to mark some points with a sawdust-lamp-black mixture. Pyramidal signals similar to those used in triangulation will also be tried out. These will be about 2.5 to 3 meters high and their frames of light aluminum tele-

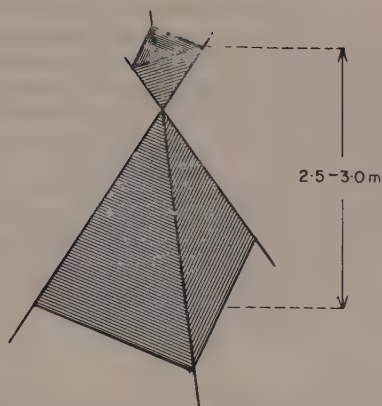


Fig. 2

scoping tubes will be covered by cloth. It is hoped that the shadowed side of the signal and the shadow cast by the signal will be depicted by the camera. From a lower altitude even the dark inside of the upper pyramid of the signal may be visible. This type of signal, if successful, may also prove to be very useful in Arctic surveying and mapping projects. The durability of these signals is, of course, a main concern.

c) Photogrammetrists well versed in glacier research should carry out field studies to ensure that important glacial features will be represented properly on the map. In particular, in field work emphasis should be placed on evidence that, when studied together with aerial photographs, will provide valuable information on the past behavior of the glacier.

5. PHOTOGRAPHIC FLIGHTS

One of the difficulties is the lack of surface definition in photographs of snow-covered areas. The situation may be critical if the photographs are made shortly after a snowfall or if illumination conditions cause strong reflections from snow surfaces.

However, if there has been no recent snowfall, the action of the wind and insolation may produce a slight roughness or waviness in snow surface, and this will make the snow surface distinct enough from the point of view of photographic imagery and photogrammetric measuring technique. Since the deciding factor is the contrast created by snowdrifts, waves, etc., low sun position at the time of the photographic flight may be very appropriate. Attention must also be paid to the exposure time. Glacier and snow surfaces should be underexposed rather than overexposed in order not to lose the details of the snow surface in the photographic image.

Debris carried by the glacier is most helpful in making the surface more identifiable, and therefore the photographic flights should be scheduled for the second half of the melting period. At that time the glacier surface is most distinct because the crevasses, moraines, and debris carried by the glacier are no longer covered by snow and are most clearly visible on aerial photographs. If it can be arranged, the time for the flight should be left to the discretion of the photogrammetrist or another member of the field team in charge of the surveying. If for any reason the decision as to the time of flight should be left to the flying crew, the field party should be warned about the forthcoming flight so that preparations may be made for a nearly simultaneous determination of control points.

In order to avoid gaps it is good practice to make at least two passes at the same low altitude, particularly if there are no reliable maps or good provisional maps over the glacier area. It is still better if two low-altitude flights could be separated by a few hours in order to provide different illumination conditions. There is a constant danger that a certain glacier area may reflect too much light towards the aerial camera because of its slope direction, and this will make the measurement and precise mapping of that particular area very difficult. With a change in the position of the sun different areas will produce glare and so each part of the glacier may be plotted from the most suitable photopairs. It is also recommended to use 80% or 90% longitudinal overlap so that there will be a sufficient choice of stereopairs for mapping.

6. MEASURING AND PLOTTING FROM AERIAL PHOTOGRAPHS

General Map of the Glacier Area

Standard photogrammetric procedures are used. If a glacier is located in unmap-ped territory, a map of only the glacier will not be satisfactory. For a more complete assessment of over-all conditions the map must include a larger area and show the surrounding mountain ranges, the accumulation area of the glacier, and the lower reaches of tributary glaciers as well. It is quite obvious that for this purpose a map on a smaller scale at, say, 1:50,000 or even 1:100,000 is satisfactory and sufficient. This map should be produced from existing aerial photographs or from the general photographic coverage of the area, and the usual rules governing topographical mapping are applied.

Glaciological Maps

Proper glaciological maps are produced at scales not smaller than 1:25,000. Even this scale imposes certain limitations in the graphical presentation of detailed glacier features and it is used mostly in the mapping of larger glaciers for which a large-scale map would have to be composed of several sheets. It is suggested that in these cases the 1:25,000 map be accompanied by detailed maps at a larger scale (at 1:10,000 or 1:5000) of the terminus and of some transverse strips at least 500 m to 1 km wide selected at characteristic zones within the glacier. The usual rules for topographical mapping are not valid for these special-purpose maps. In particular, the usual standards of accuracy for contour lines must be replaced by more rigorous ones. In other words, the accuracy of contour lines must be divorced from the scales of plotting—and most definitely from the contour interval used—and it should represent the maximum accuracy obtainable from the given aerial photographs. Personal errors and errors inherent in current topographical plotting must be eliminated as far as possible by special precautions and by special plotting procedures. Continuous contouring must be replaced by point-by-point contouring supplemented by spot elevation readings if necessary. In particularly critical zones the contouring must be performed twice and the mean contour line accepted as the final one. Any spot elevations must be determined with the utmost care. It is best if the operator can record the elevations automatically in several pointings without being cognizant of individual readings. The accuracy of contours and of spot elevations should be stated on the map.

The horizontal accuracy is less critical but equally important. Plotting table errors should not exceed 0.1 mm, which is of a magnitude characteristic of most first-order plotters. Extremely high accuracy requirements, sometimes excessive flying

heights which must be used for reasons discussed previously, and plotting difficulties of snow-covered surfaces, etc., make the use of first-order instruments in the plotting of glacier maps mandatory.

Apart from special accuracy requirements, the glacier map must contain many details of glacial morphology which would not be included in an ordinary topographical map. This is another reason why the scale of glaciological maps must be relatively large.

Of course it is obvious that the above requirements are restricted to the measuring and mapping of glaciers and their bordering bands of solid terrain to which glacier measurements are tied in. For the remaining part of the map, the standards accepted in topographical mapping are applicable and, in particular, areas other than glaciers can be plotted from different high-altitude photographs.

Special Measurements

Aerial photographs may also be used for special measurements, such as for profiling glaciers, for determining the rate of flow, for producing comparative plots from earlier photographs, etc. Some of these applications of aerial photogrammetry may call for re-photography at specific time intervals and may give rise to further signaling problems. To determine how practical and how justifiable the use of aerial photogrammetry can be in these particular applications will be an aim for future investigations in glacier research.

N.R.C. No 6236

ON THE QUESTION OF THE RESEARCHES OF THE GLACIERS BY THE METHODS OF THE ELECTRICAL PROSPECT

B.A. BOROVINSKY (U.R.S.S.)

Glaciological group of the Geographical Department of the
Academy of Sciences of the Kazakh S.S.R.

SUMMARY

1. The Academy of Sciences of the Kazakh S. S. R. has researched the glaciers in Kazakhstan by the electrical prospect methods, which have the following advantages in comparison with the seismic and other different research methods:

versatility, varieties of problems that cannot be solved by other methods, better adaptability to the real conditions and sufficient precision of the results received.

2. Still, wide use of the methods of the electrical prospect under the conditions of the glaciers is difficult.

There are no clear recommendations for the field work methods and there are no necessary apparatus in serial production.

3. For defining the thickness of the glacier by methods of electrical prospect dipol-radial soundings and soundings of the aggregat setting are recommended.

In this case the dipol setting can be used successfully for the research of the big tickness under the conditions of horizontal electrical heterogeneity, which takes place in the middle part of the tongue of a glacier and in the province of circus.

The combinative setting can be used successfully for the ice thickness research in the province of thawing of the glacier.

4. While interpreting the material, two assumptions can be made depending on the place of the work on the fase of the theorem about unambiguous effect isometric volanes.

a) The research limits are supposed to be horizontal and plane paralell. The circuses of the glaciers satisfy with these conditions roughly.

b) The body of the glacier can be accepted as half a cylinder, stretched along its axes.

The ideal condition in this case is the middle part of the length of the tongue.

Interpretation for the case «A» is realized by paletka of usual soundings; in the second case by paletka of the lateral carottag sounding.

The theoretica. crookedes transform with interpretation crookedes dipol-radial soundings of the formule

$$\varrho r = \varrho_0 \left\{ 1 - \frac{1}{4\varepsilon} [\varrho(r + \varepsilon) - \varrho(r - \varepsilon)] \right\}$$

Where

ϱ — is apparent specific resistance for dipol-radial soundings

ϱ_0 — is for K S simmetrical sounding

ε — is some interval, convenient for calculations.

c) The interpretation of electrical prospect material is examined. According to the results of the seismic work on the glacier Tuyuksu and Shumsei the coincidences of ultimate results is good.

RÉSUMÉ

1. L'Académie des Sciences de la R. S. S. de Kazakhie a entrepris des travaux sur l'étude des glaciers du Kazakhstan par des méthodes de prospection électrique qui possèdent une série d'avantages comparées aux méthodes de recherches sismiques et autres : souplesse, variété des problèmes impossibles à résoudre par d'autres méthodes, une meilleure accommodation à la situation ambiante, une précision suffisantes des résultats obtenus.

2. L'emploi à une large échelle des méthodes de prospection électrique est toutefois difficile à réaliser dans les conditions des glaciers. On manque de recommandations claires sur la méthode des travaux sur place et les interprétations, les appareils nécessaires à l'entreprise des travaux ne sont pas encore fabriquées en série.

3. Afin de déterminer l'épaisseur de la couche de glace par les méthodes de prospection électrique il est recommandé de se servir de sondages radiales dipôles et de sondages à l'aide d'une installation combinée. L'installation dipole peut être utilisée

avec succès au cours des recherches sur les couches épaisses dans les conditions d'hétérogénéité électrique horizontale qu'on rencontre dans la partie médiane de la langue du glacier et dans la région du cirque. L'installation combinée peut être utilisée avec succès pour déterminer l'épaisseur de la couche de glace des régions de fusion du glacier.

4. Lors de l'interprétation de la documentation obtenue, deux admissions sont possibles en fonction de l'emplacement des travaux et en vertu des théorèmes sur l'égalité du signe de l'effet de volumes de grandeur égale :

a) les limites étudiées sont considérées comme horizontales et parallèlement planes. Les cirques des glaciers répondent approximativement à ces conditions;

b) le corps du glacier peut être considéré comme étant un demi cylindre allongé le long de son axe. La partie de la langue de largeur moyenne serait, dans le cas envisagé la condition idéale.

L'interprétation pour le cas « a » est réalisée d'après les palettes de sondage ordinaire et dans le second cas — d'après les palettes de sondage latéral par carottage. Lors de l'interprétation des courbes de sondage dipôles et radiales, les courbes théoriques se transforment d'après la formule :

$$\varrho_r = \varrho_\theta \left\{ 1 - \frac{1}{4\varepsilon} [\varrho(r + \varepsilon) - \varrho(r - \varepsilon)] \right\}$$

où

ϱ_r — la résistance spécifique apparente pour le sondage dipole-radial,

ϱ_θ — pour résistance apparente du sondage symétrique,

ε — un certain intervalle désirable lors des calculs.

5. L'interprétation de la documentation obtenue par la prospection électrique a été vérifiée d'après les données des travaux sismiques sur les glaciers Toujousou et Choumski.

La concordance des résultats finaux est satisfaisante.

During the period of the I. G. Y.-I. G. C. — 1957-59 the Geographical Department of the Academy of Sciences of the Kazakh SSR carried out glaciological researches in the high-mountained regions of the Zailiysky and Djungar Alatau. A great volume of this work has been accomplished by methods of applied geophysics, which included seismic exploration, electrical exploration and magnetometry. Principally, these researches bore an experimental character. As the practice of the work in the mountains of the Zailiysky Alatau had shown, the electrometric methods have a number of advantages over the seismic and other methods. These are the small weight and dimensions of the electrometric apparatus, simplicity of operation, sufficient precision of measurements and the obtained results and, lastly, the large number of problems that are being solved. The methods of electrical exploration can successfully be applied on glaciers for the solution of the following problems:

1. Determination of ice thickness.
2. Measurement of the motion speed of inner ice layers.
3. Determination of the inner structure of a glacier and the elastic constances of ice by the measurement results of the electrical parameters.
4. To ascertain sections of increased thawing of the ice down the glacier's surface.
5. Measurements of the inclination angle of glacier borders.
6. The study of the inner structure and the hydrological peculiarities of moraines.

The electrometric work on ice was begun a comparatively short time ago. Therefore, there are no sufficiently elaborated methods of the field work and the interpretation of the obtained material is not yet available. The necessary special standard apparatus are not yet taken up into serial production. By studying the structure and thickness of the glaciers by methods of electrical explorations, combined soundings and soundings of dipole-radial sets were used. Combined or counter soundings are called soundings carried out by sets AMNB∞, ∞MNB and usual symmetrical sounding with AMNB, that is being carried out simultaneously. By such sounding a mark SR for the set MNB is being made at first, then follows the mark SR for the set AMN, the third mark is for the usual symmetrical set AMNB and is the controlling set

to the former two, because the SR of the third mark ought to equal to half of the sum of the former two marks.

If there is no horizontal discontinuities in the layers under study and the thicknesses of the layers are constant, there can be no difference in the SR of the tripol sets. In carrying out electrical exploration work on a glacier, it often becomes necessary to undertake soundings in places where the thickness of ice changes rapidly. Besides that, owing to genetical and even exposition peculiarities of the researched glacier region, there may be cases of noticeable change of the composition of the ice (inclosure of a terrigenous fraction, change of temperature, porosity and filtration features in the local region of the glacier surface) and therefore changes in the specific electrical resistance. If the similar curves are even slightly dislocated in relation to each other, in this case there exists a horizontal homogenousness of composition and an approximate constancy of the thickness.

Usually the sounding curves are always interpreted as having been taken over the horizontal homogenous layer (a case of practical absence of horizontal gradient of specific resistance). Here it is assumed that the observed layers are bedded horizon-

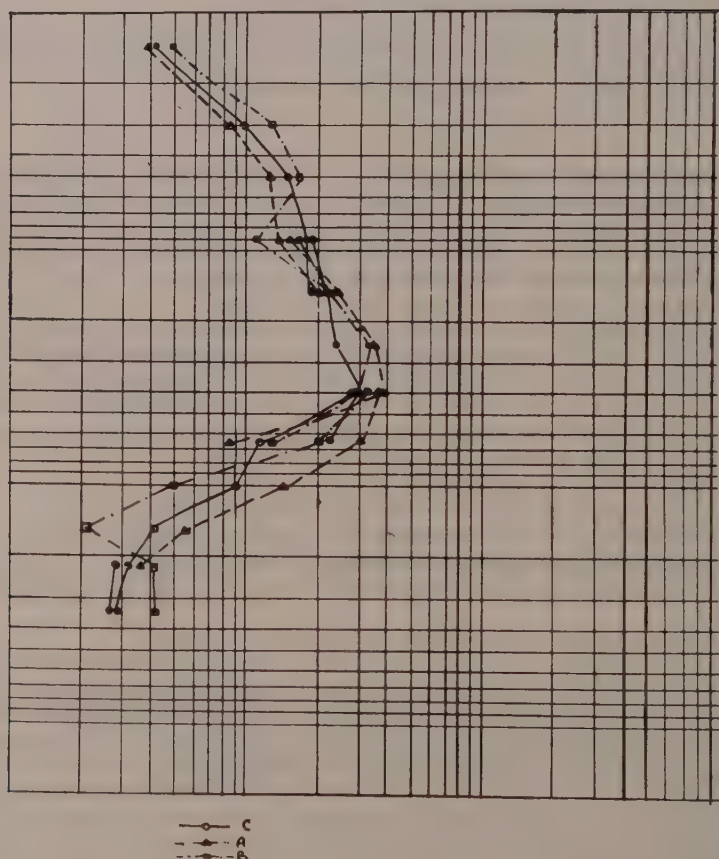


Fig. 1 — Typical sounding curves by combined set, obtained on the glacier Molodeshny.

tally and if there is no data available about the angle and azimuthal fall of the surface of the section, such assumption often leads to a number of wrong conclusions when are interpreted the curves.

There is paletka available for the interpretation of curves, that were received above an inclined limit of the section by arranging of the electrode, parallel and perpendicular to the contact line of the stratum.

In view of the fact that the border of the bed of the glacier may be considered as the inclined limit of the section, by establishing the soundings not on the axis of the glacier, but near its border, the calculation of the incline by the interpretation of the soundings may be defined by the above mentioned paletka. However, the use of such paletka presents difficulties on account of the absence of any information regarding value of the inclination angle of the glacier border. When working on the glaciers there often occur cases of low inclination angles of the glacier bed and of inconstancy of resistance of the ice as the function of depth. In such cases usual soundings with one azimuth of arrangement of electrodes is inapplicable. The determination of the horizontal dissimilarity is most fully achieved by multiazimuth soundings with the

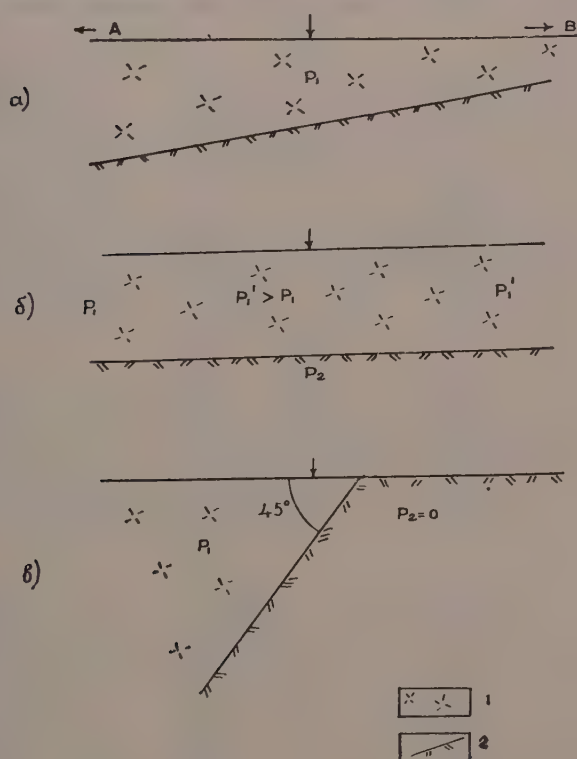


Fig. 2 — Schematic View of two-layer sections with horizontal dissimilarity.
a) Inclined limit with small sinking angle.
b) The ice resistance changes smoothly along the section.
g) Inclined limit with sinking angle of 45° .
Conventional sign: 1 ice
2 bed stratum.

combined set. As a rule, the arrangement of the electrodes are accomplished both along and across the glacier.

When the curves SR AMN, SR MNB and SG AMN B are near to each other, which speaks in favour of the presence of approximate homogenousness of the layers of the observed part, the palette can be utilized for the horizontal layers. For instance, the glacier Molodeshny, situated in the Tuyuksu glacier group, by virtue of its genetic peculiarities, has an almost horizontal structure of the bed. in the middle part The sides of the bed are not distinctly enough formed by the glacier, and therefore, when interpreting the WS, the theoretical curves for the horizontal layers fully fitted here. F.1. In this case the influence of the uneven relief of the bedding stratum tells mainly by arrangement of the electrodes that are placed near the border of the glacier, and, as a rule, will tell very little on the arrangement of the electrodes that is directed to the opposite side.

The surface of the valley glacier that is comparatively slightly dislocated, practically does not exert any influence upon the sounding curves. By consideration of the theoretical curves (Fig. 2a,b,c) for the case of an inclined bi-layered cross section with a direct or alternative resistance (Fig. 3a,b,c), it can be concluded the change of resistance of the first layer by stretching at constant thickness will not influence

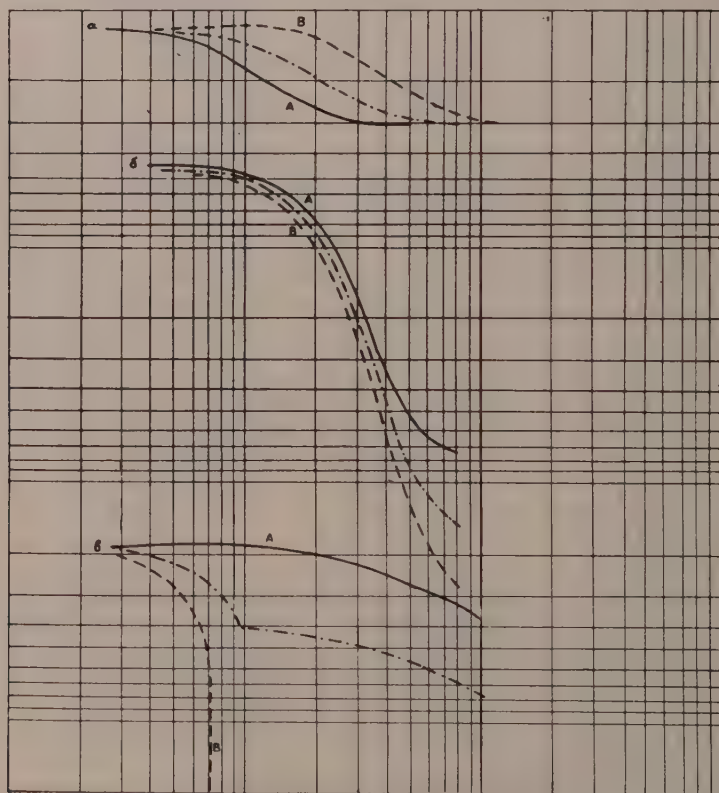


Fig. 3 — Calculated curves of soundings with combined set above the sections shown on the former graphic.

strongly the results of the interpretation, if the resistance changes little within the distances that are placed near the feeding of the electrodes.

The dipole soundings have some advantages over the soundings with the combined set. However, the curves of the dipole soundings are very sensitive to the dissimilarity of the upper layers.

By crossing with the dipole the vertical or inclined contact (on the glacier where cracks with moistened walls will be, on the moraine-pools, filled with small terrigenous fraction of a good electrical conductivity) the curve of the dipole sounding suffers a rupture which is not noticed at sets with three or four electrodes. Apparently for the determination of comparatively small depths, as, for instance, the research of glaciers of a shallow depths in the region of the tongue, it is convenient and advisable to use the combined set, but for glaciers of large depths—the dipole set.

Soundings on the glacier with several azimuths by means of combined set (across and along the glacier) with a verification by means of a four electrodes set, and also sounding with a dipole set were carried out on the Tuyuksu glacier in 1957.

Then, in 1958 and 1959 on the glaciers Molodeshny, Igly Tuyuksu, Zoya Kosmodemyanskaya and Manshuk Mametova such soundings, were carried out solely, because usual soundings could not fully answer all questions due to distortions of influences caused by:

1) Unevenness of relief, 2) dissimilarity of ice in the horizontal direction, 3) unconstant thickness of the ice, bad and in the course of time changeable grounding resistances. The occurring dissimilarities are easily ascertained by carrying out of several soundings on a straight line according to which, the cross section is being composed. Interpretation is considered chiefly from the middle curve that was received with the aid of the symmetrical set. The other curves are calculated for the purpose of ascertaining dissimilarities and also of the general inclination of the glacier bed.

The curves for sets with three electrodes are interpreted independently of the rest, and the depths of the repository of the section, received by the interpretations, are carried over to the points that are away from the centre of the set to the side of the electrode arrangement, at a distance of a double repository depth of the section limit.

Thus, out of each combined sounding for each section limit three points can be received. The received angle of inclination of the limit is being conformed with that of the neighbouring soundings. If the angles of inclination do not conform, there exists a variation of the specific electric resistances of the layers. For the interpretation of the curves of the dipole-radial soundings, the theoretical curves of usual symmetric set were transformed by formula.

$$Pr = P\theta \left\{ 1 - \left[\frac{1}{4\varepsilon} (p + \varepsilon) - (p - \varepsilon) \right] \right\}$$

Where Pr —the seemed resistance for dipole-radiation set.

$P\theta$ —the seemed resistance for the symmetric set.

ε —Some interval, that is convenient for counting over, the value of which is

determined by formule $\varepsilon = \frac{M}{2} la$ where M = module of the paletka.

Dependent on the place where the electrical soundings are carried out and on the character and type of glacier, the interpretation of three methods had been applied.

1. Interpretations in the case of horizontal bedding of the layers were applied in the lower part of the glacier Molodeshny and also in the region of the firn basin of other glaciers; the electrodes that are feeding the lines are comparatively far from the borders, and the borders of the glacier by visual observation were not less than at a five times the approximate thickness of the ice.

2. For the case of one inclined limit of the section. This case is frequently being met by carrying out soundings near the border of the glacier by arranging the feeding electrodes parallel to the extent of the glacier.

The interpretation was complicated because the very angle of inclination was unknown and it had been carried by usual theoretical or specially built paletka.

3. Interpretations of soundings that had been carried out in the central part of the glacier on an axis by the approximate consideration of its bed as a semi-cylinder. The interpretation for the 3-rd case has been carried out by the paletka of the side carottage-sounding. At first, by the paletka of the type V, the values of the apparent resistance of the upper and lower layers and the approximate values of their depths were determined. Then, by the paletka—the gradient of the sounding or by the speciale rebuilt paletka for dipol sounding, definite results concerning the radius of the unknown semi-cylinder had been obtained. By the determination of the thickness of the ice there proceeded from conjecture that equal volumes with equal physical parameters at considerable distances cause equal anomalous effects. Hence, it follows that if in our case the section of the glacier, in the place that it is being studied, deflected from the semi-cylinder section, the area of this section equalled the area of the semi-cylinder with the radius of the section.

Lastly: we measure the apparent resistance of the region through which an electric current is passed. This resistance occurs as an average and its significance is justified for a certain entirely definite volume. The properties of the strata are equal. The seemed resistances are received, proceeding from the summary significance of the resistances on the entire area, it becomes clear that the effects of equal volumes filled matter of equal properties will be equal.

The seismic exploration that was carried out on the glacier Tuyuksu (Zailiy Alatau and Shumsky (Djungar Alatau) did not allow to single out the dissimilarity of the structure of the glacier body. The reason for it was the comparatively small thicknesses of the glacier as such and insignificant changes of the elastic ice constants.

The rectilinearity of *hodographs* of the longitudinal waves and the parallelism, of the *hodographs* that are overtaking the rectilinearity received on these glaciers, prove the absence of a vertical gradient of speed in the thick of the ice. However, according to the data of the electroexploration and of the structure analysis, this conclusion is conditioned by insufficient precision of the seismic method. Electrical soundings, as a rule, bring out a two-layer structure of the ice: the upper layer, the electrical resistance of which changes by the time and depends on the influence of climatic conditions, and the *lower layer*- with its more constant characteristics. Besides that, the accumulation of the firn in the accumulation region is being quite clearly demarcated from the below-bedded depth ice. If in the ablation region, without considering the zone of the intra season and season fluctuation of the temperature, the soundings can be interpreted by two-layer paletka ice-stratum, so higher up in the region of the firn line and the firn, they are interpreted exclusively by the three-layer paletka, by taking into consideration the accumulation of the firn.

The thickness of the ice of the glacier Zentralny Tuyuksu has been determined by electrometric and seismic methods. The disposition of the transverse profiles on these glaciers are described in the report of the snow cover, that is included in the present collection.

In the region of the first profile the thickness of the ice was determined by the seismic method equalling to 37 m, and by the electrometric method — 33 m, whereby the curve of the electrical sounding shows thicknesses of the upper zone equalling to 5-6 m. The relation of resistances of the lower layer to the upper W2 equals to 1.9. The maximum of the apparent resistance equals approximately 0.7 actually. The thickness of the ice on the third profile, according to the seismic and electrometric data, equals to 53-54 m. The angle of inclination of the bed towards the horizon equals 6°. The depth of the first layer equals to 6-8 m.

The thickness on the fourth profile, determined by the seismic method, equals to 50 m, by electrometry—59 m. The depth of the layer of intraseason and season changes reached also 6-8 m.

The thickness on the fifth profile, after a correction of the received semi-cylinder radius had been made, equals to 89 m. The seismic method showed 74 m, and the methods Lagalli and of balances (N. Palgov)—86 and 87 m. On the sixth profile the thickness of the ice was determined by electrical exploration to be 135 m. The sounding shows clearly a dissimilation, in this place at the depth of 10-12 m probably caused by the inclosure of terrigenous fraction.

On the seventh profile of a general ice thickness of 115 m is clearly shown an upper layer, which is characterized by icy-firn-streaks of a general thickness of 21-23 m. The sounding curve has in its left asymptote an unsettled saw-shaped form, probably caused by the alternation of the above mentioned layers.

The comparison of data of the structural analysis and the distribution of temperatures with depth in the ice had shown, that the electrical properties are closely interconnected with the structure of the ice and the temperature regime of the glacier.

The intraseason and the season zone of the temperature fluctuations (tropozone) is distinguished by electrical properties on the sounding curves, probably, because the direct and inversal temperature stratification causes, to a certain extent, the change of the dimensions, structure or orientation of the ice crystals. The lower bedded stratozone is characterized by a direct temperature stratification only. It gradually passes into the zone of relatively constant temperatures. Therefore there are no sharp temperature contrastings and, if, nevertheless, there is a change of the structure, it occurs smoothly, under the influence of the regular change of pressure, temperature and the motion speed within the thickness of the ice. It may also be surmised, the existence of a connection between the electrical and elastic properties of the ice of mountain glaciers, as the mentioned properties depend on the porosity, density and mineralization of the film ice water.

Summarizing the following conclusions may be drawn:

1. The research of glaciers of the middle latitudes with the aid electrometry of gives positive results.
2. Combined and dipol sets occur most applicable and advisable.
3. By comparison of the data of electrical soundings with the of observations results of the temperature regime and the structural analysis of the ice, a dependence of the electrical conductivity on the thermoconductivity, conduction of temperature and structure of ice, may be surmised.
4. The correctness of the interpretation depends on the extent of consideration of the influence of the glacier borders and the dissimilarity in the horizontal direction. The interpretation of the sounding consists of three stages:
 - a) determination of the inner resistance of the layers;
 - b) ascertainment of the area of the profile section by equivalent reality;
 - c) determination of the thickness of ice.
5. The main difficulty by performing work on the ice consists of the fact that the groundings as such present a great resistance of several tens and hundreds of megometers. Whereby, usual galvanometers cannot be used and it is necessary to apply electronic apparatus, a most stable scheme of which has not yet been elaborated.
6. The paletka, that are being used on usual mountain strata cannot always be adapted for the research of glaciers. Some paletic material, obtained by the author, will be published in the near future.

The problems stipulated at the head of this report, that are solved by electrical exploration methods, had been applied on the Tuyuksu glaciers in the Zailiysky Alatau. However, we could not fully elucidate them within the limited size of this report. The results of this work were partially published before, another part of it will be published in the near future.

GEOPHYSICAL SURVEYS ON GILMAN GLACIER, NORTHERN ILLEMERE ISLAND

J.R. WEBER ^[1], N. SANDSTROM ^[2] and K.C. ARNOLD ^[3]

SUMMARY

During the summers of 1957 and 1958, geophysical surveys were carried out on Gilman Glacier and the adjoining ice cap of northern Ellesmere Island. Seismic refraction and reflection measurements showed that a maximum compressional wave velocity of 3795 m/sec was reached at a depth of 50 m on Gilman Glacier. On the ice cap a maximum velocity of 3810 m/sec was reached at a depth of 100 m. Reflection profiles were calculated at various sites. On the glacier, the ice thickness varied from 380 m, 5 km from the terminus, to 760 m, 19 km from the terminus. On the ice cap, the ice thickness varied from 400 to 800 m. The mean shear stress at the bed of the glacier was found to be 0.85 bars. Calculations of the rate of discharge through a cross-section indicates that Gilman Glacier has a slightly negative regime at the present time.

During the summer of 1958, more than 200 gravity stations were established over the area of the seismic survey. The regional Bouguer anomaly was calculated from the known ice thickness at a few selected sites along the seismic profiles, and was then extrapolated for the whole area. The ice thickness was then calculated from the gravity measurements with assumed densities of 0.9 and 2.71 g. cm⁻³ for ice and bedrock respectively. Agreement between bedrock profiles as determined seismically and gravimetrically was very close.

Measurements of surface movement of Gilman Glacier indicated a maximum velocity of 25 m/year.

RÉSUMÉ

Au cours des étés de 1957 et de 1958, certaines expériences géophysiques furent effectuées sur le Glacier Gilman de même que sur la calotte glaciaire attenante, situées dans le secteur septentrional de l'île Ellesmere. Les mesures de réfraction et de réflexion sismiques prises à l'intérieur du glacier Gilman établirent la vitesse maximum de l'onde de compression à 3795 m/sec à une profondeur de 50 m, tandis qu'à 100 m sous la calotte la vitesse maximum était de 3810 m/sec. Plusieurs profils de réflexion furent dressés en divers points de la région. Sur le glacier, l'épaisseur de la glace était de 380 m à 5 km du terminus, et de 760 m à 19 km du même endroit. Sur la calotte glaciaire, on a noté des épaisseurs allant de 400 à 800 m. La force moyenne de cisaillement du lit du glacier a été évaluée à 0.85 bars. De plus les calculs effectués transversalement au glacier Gilman, touchant son rythme d'ablation ont démontré que son régime actuel était quelque peu négatif.

Durant l'été de 1958, plus de 200 stations gravimétriques étaient en opération dans la région affectée aux expériences sismiques. L'anomalie régionale de Bouguer a été calculée à partir de certaines données connues sur l'épaisseur de la nappe de glace le long des divers profils sismiques, et extrapolée par la suite pour toute la région. On procéda, de plus, au calcul de l'épaisseur de la glace d'après certaines mesures gravimétriques, en assumant au préalable à la glace et à la roche des densités de 0.9 et de 2.71 g.cm⁻³ respectivement. La confrontation des profils du socle rocheux établis d'après les méthodes sismiques et gravimétriques démontrèrent une très grande similarité.

Les mensurations sur le mouvement superficiel du glacier Gilman ont évalué sa vitesse maximum annuelle à 25 m.

[1] Gravity Division, Dominion Observatory, Department of Mines and Technical Surveys, Ottawa; formerly of the Department of Physics, University of Alberta.

[2] Scripps Institution of Oceanography, University of California, La Jolla; formerly of the Geophysical Laboratory, Department of Physics, University of Toronto.

[3] Geographical Branch, Department of Mines and Technical Surveys, Ottawa; formerly employed by the Department of Physics, University of Toronto, and by the Department of Geography, McGill University.

INTRODUCTION

The surveys described in this paper were carried out on the Defence Research Board's Operation "Hazen", the Canadian IGY expedition to the Lake Hazen area of northern Ellesmere Island, 1957-58; they covered Gilman Glacier and the ice cap in the vicinity of Mount Oxford (Fig. 1). In 1957, six reflection profiles were obtained



Fig. 1 — Map of Gilman Glacier, northern Ellesmere Island to illustrate geophysical surveys. (Prepared by the Geographical Branch, Department of Mines and Technical Surveys, Ottawa. Form lines from surveys by K.C. Arnold on Operation «Hazen»).

across Gilman Glacier and the névé at its head (Figs. 2 and 3); in 1958, a refraction profile was made on Gilman Glacier, and four refraction and reflection profiles on the ice cap. A gravity survey was undertaken over the same area in 1958, in order to determine how far depth measurements by the gravity method agreed with seismic measurements, and to supplement the data obtained from the seismic survey (Fig. 1)

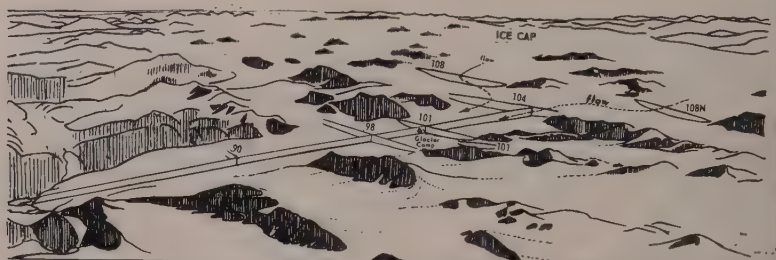


Fig. 2 — Block diagram of Gilman Glacier from the east. (From a drawing by G. Falconer).

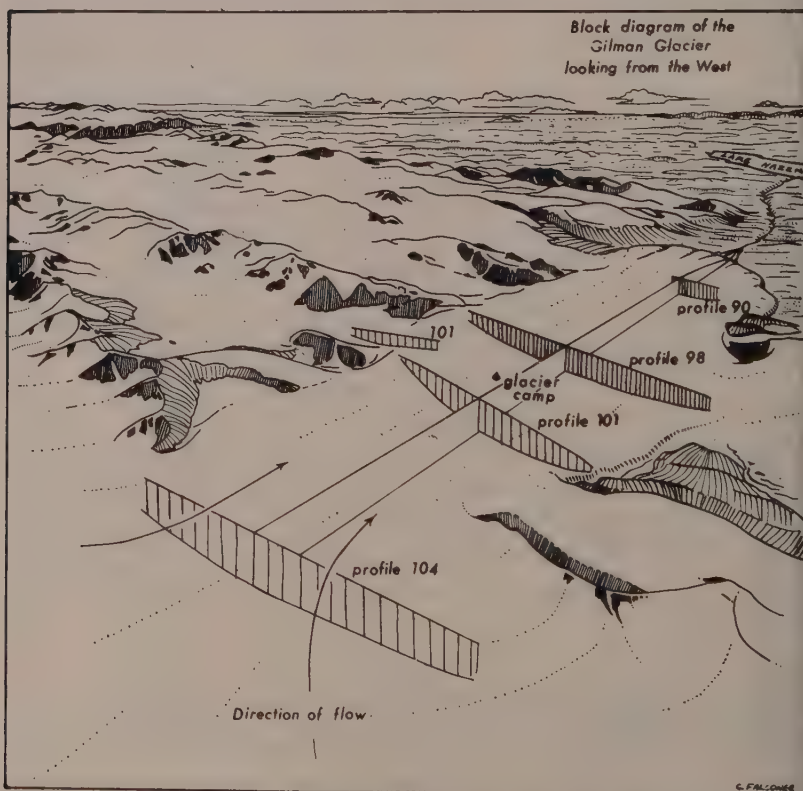


Fig. 3 — Block diagram of Gilman Glacier from the west. (From a drawing by G. Falconer).

A local network of more than 200 gravity stations was established. A triangulation and levelling survey was undertaken to establish the positions and elevations of seismic and gravity stations, and periodic observations were made on a network of stakes planted in the glacier to measure surface movement (Fig. 1).

2. SEISMIC SURVEY

2.1. *Equipment and field procedure*

The seismic party was equipped with a high resolution seismograph built by Houston Technical Laboratories. Communication between recording and blasting equipment was by telephone, and time signals were transmitted by time-break line, owing to the failure of two radio transmitter-receivers. Thus, work was restricted to distances controlled by the amount of telephone cable available.

For refraction measurements, the geophone spread and shot points were laid out in a straight line. For reflection measurements, the geophone spread was laid out in an *L*-pattern with shot points at measured distances from the centre of the geophone spread and in line with the branches of the *L*. All distances were measured with a 100-metre steel tape and all vertical angles with a Wild *T2* theodolite. Thirty-six geophones were connected to the spread in groups of three spaced equally along the length of the cable. The available geophones were all of vertical type and therefore most sensitive to longitudinal motion.

2.2. *Records*

A total of nearly three hundred records was obtained. The records from Gilman Glacier generally exhibit quite clearly marked first arrivals and p_1 -reflections. *Sp*-reflections appear on some of the records and a few *S*-reflections are discernible. Records from profiles 108 and 108 N (Fig. 2) show similar characteristics. The ice cap records are more complicated; besides first arrivals, they show surface-reflected *p*-waves on almost all records at distances greater than 400 m, bottom-reflected *P*-waves, and *Sp*-waves.

2.3. *Analysis of records*

From travel-time relations the velocity-depth curves were constructed from the equation:

$$z(v) = \frac{1}{\pi} \int_0^{\Delta(v)} \cosh^{-1} \frac{v}{C(\Delta)} d\Delta,$$

where z = depth; v = horizontal velocity (variable with depth); Δ = variable distance between shot point and receiver; $\Delta(v)$ = distance, where the velocity $\frac{d\Delta}{dt} = C(\Delta)$ takes the value v ; t = travel time; and C is a constant. Implicit in this relation is the assumption that the velocity does not decrease with depth anywhere between the surface and the depth z .

The velocity-depth curves for three different areas are shown in Fig. 4. For comparison two curves obtained by Bentley and others⁽¹⁾ and by Joset and Holtzscherer⁽²⁾ on the Greenland ice cap are also shown. The differences between

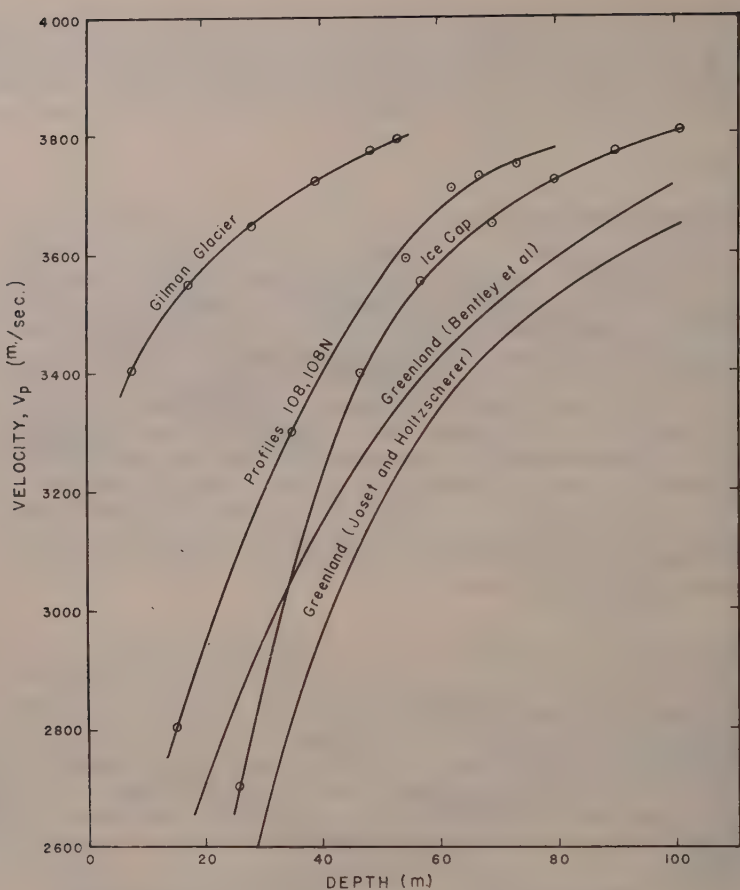


Fig. 4 — Velocity-depth curves from first arrivals.

the curves can mainly be attributed to varying rates of accumulation and ablation and of compaction of the firn where present since the difference are most marked in the near-surface layers. The greatest difference is seen to exist between conditions on the glacier, where ablation predominates, and conditions on the ice cap, where a layer of firn overlies the ice. Fig. 4 shows that if a layer of snow and firn approximately 50 m in thickness were removed from the ice cap the velocity-depth curve for the ice cap would be similar to that for the glacier. At depths below the firn the velocities appear to be mainly governed by such factors as temperature and pressure.

The maximum velocity in glacier ice was found to be 3795 ± 10 m/sec., which was reached at a depth of about 50 m and corresponded to a shot distance of about 600 m. At distances greater than 600 m the travel-time curve is straight and gives a velocity of 3795 ± 10 m/sec.. The maximum velocity obtained on the ice cap was 3810 ± 10 m/sec., which was reached at a depth of about 100 m and corresponded to a shot distance of about 1000 m. There is some uncertainty about the depths and distances quoted owing to the very slight curvature of the travel-time curve.

2.4. Surface-reflected waves

The ice cap records for shot distances greater than 400 m show multiple, surface-reflected p -waves as a series of pulses of almost identical form. It was found that these could be analyzed as arrivals which had been reflected once, twice, etc. at the surface: i.e. an arrival which had been reflected once at the surface had a travel time twice that of the direct p -wave at half the distance, while the doubly reflected wave had a travel time three times that of the direct p -wave at one-third the distance, etc. This surface "channelling" effect was occasionally so appreciable as to mask other weaker arrivals, making interpretation difficult and occasionally impossible⁽³⁾.

2.5. Bottom-reflected waves

Three main types of bottom-reflected waves were recorded: reflected p -waves (Rp), waves converted from S to P upon reflection (Rsp), and S -waves (Rs). The P -waves in a few cases were reflected twice.

If the ray paths of the reflections are not greatly curved, the slope of a plot of Rp^2 against Δ^2 is $\frac{1}{\bar{V}_p^2}$, where \bar{V}_p is the average velocity for a vertically travelling

wave. For glacier ice the velocity as determined by the least squares method was 3769 ± 5 m/sec. The error as given represents the standard error determined only by the scatter of points, and does not represent uncertainty in velocity determinations.

In a consideration of the reflection measurements made on the ice cap, a complication arises due to the fact that the ray paths can no longer be considered as straight, since the velocity changes very rapidly in the surface layers. To reduce to the situation where Rp^2 against Δ^2 could be plotted, travel time between surface and a depth of 100 m (i.e. the time taken to traverse the region of rapidly changing velocity) was subtracted from Rp and the corresponding distance from Δ . The average vertical velocity determined in this way was 3771 ± 3 m/sec for ice below a depth of 100 m.

No similar analysis could be carried out for the S -waves, but for calculation of reflection points it was assumed that the average shear wave velocity was given by $\bar{V}_s = \frac{1}{2} \bar{V}_p$.

For reflection profiles below the firn line, the reflecting horizon was determined by making a three-dimensional analysis of reflected arrivals along the L -shaped spread⁽⁴⁾. The various steps in computation were programmed and performed by "Ferut", the electronic computer at the University of Toronto Computation Centre. This procedure, however, was not applicable in cases where the ray paths were considerably curved, and the data therefore insensitive to directional analysis. Thus, for the ice cap data the assumption was made that reflection occurred in a vertical plane midway between shotpoint and detector. This assumption becomes questionable when the bottom topography is undulating, and in this case corrections should be made after a first approximation for the shape of the bottom has been obtained. On Gilman Glacier a first approximation gave reasonably good indications of the topographic features beneath the ice.

The longitudinal profile of Gilman Glacier is shown in Fig. 5, constructed from information obtained by surface surveying⁽⁵⁾ and by seismic and gravitational depth determinations. The bottom of the glacier appears rather uniform, with a gentle down-glacier slope averaging about 0.65 per cent for a distance of about 15 km, compared with a slope of 2.5 per cent at the glacier surface. The transverse sections are regular in shape and the normal cross-sections have a wide and flattened U-shape. One of the transverse sections is shown in Fig. 6. Similar profiles have been reported from one of the outlet glaciers of the Penny Ice Cap in Baffin Island⁽⁶⁾.

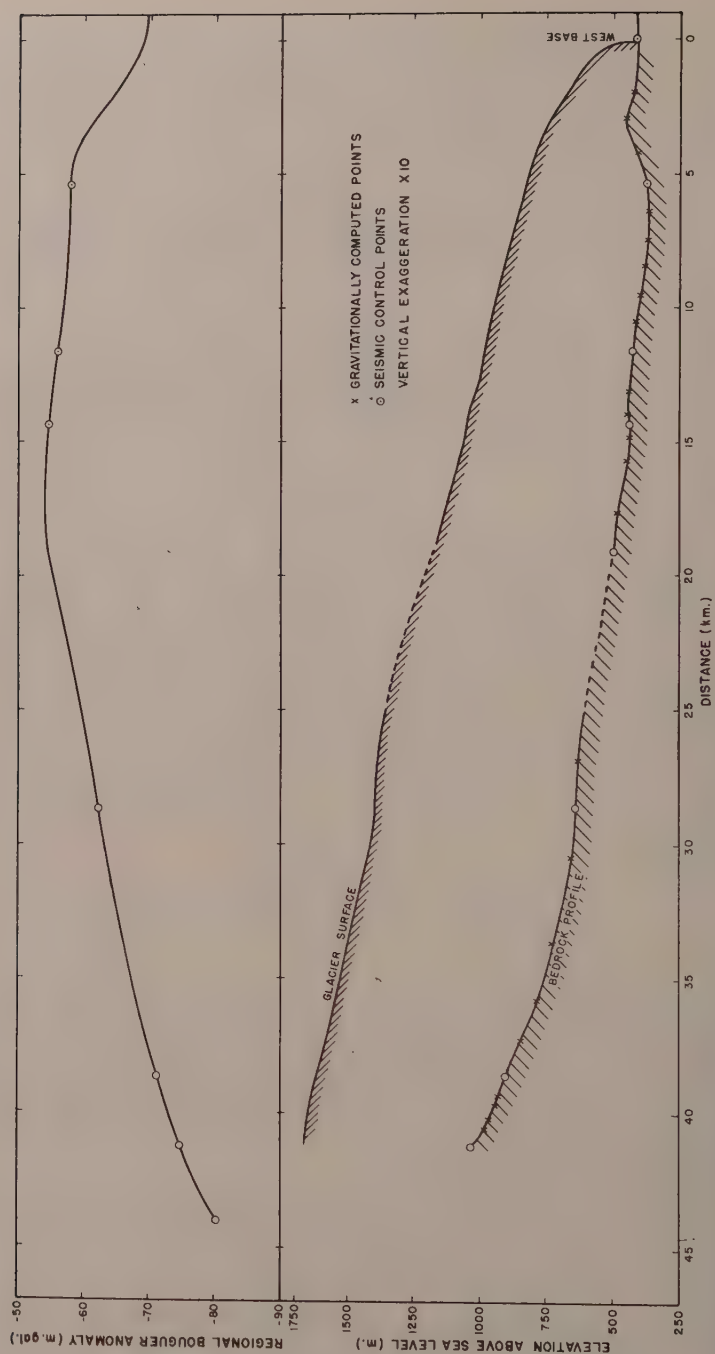


Fig. 5 — Regional Bouguer anomaly and longitudinal profile, Gilman Glacier

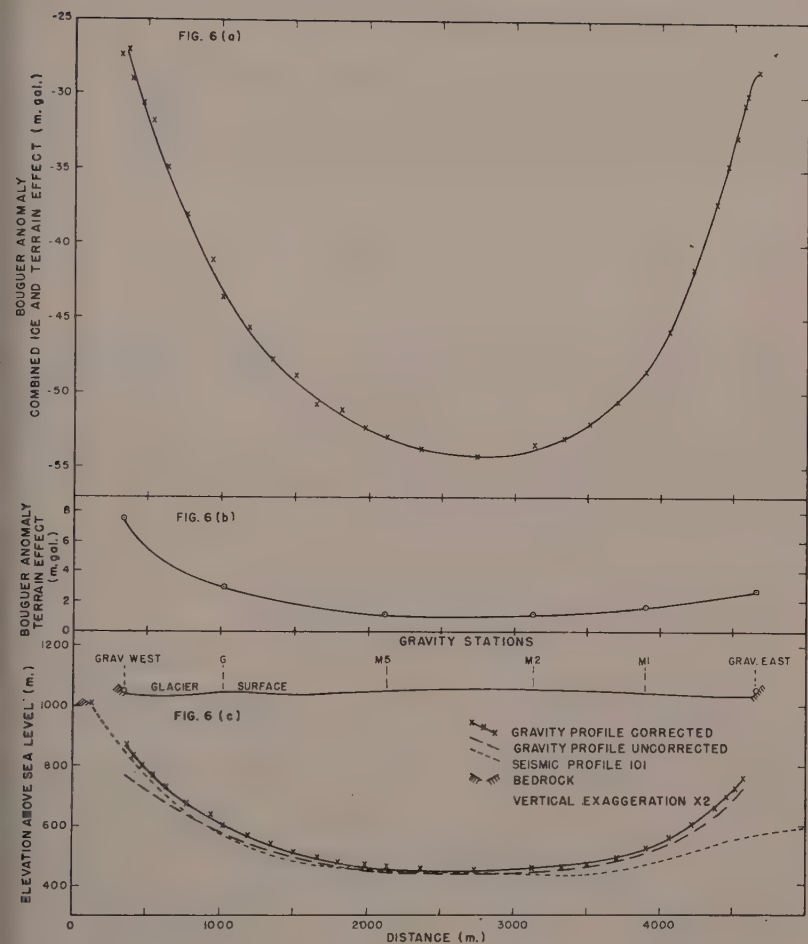


Fig. 6 — Transverse profiles and Bouguer anomaly, Gilman Glacier.

2.6. Discussion of results

The transverse glacial sections, obtained as described above, can be used to calculate the average shear stress at the glacier bed. In a glacier of constant cross-section flowing uniformly, the average shear stress (τ_{av}) at the bed is given by simple resolution of forces (⁷). Thus

$$\tau_{av} = \rho g \frac{A}{p} \sin \alpha$$

where ρ is the average ice density, g the acceleration due to gravity, A the area of cross-section perpendicular to the bed, p the perimeter, and α the slope of the surface.

Values of A and p were measured for three profiles across Gilman Glacier (Figs. 2 and 3), and the results are tabulated below. Values of 0.90 g cm^{-3} and 985 cm sec.^{-2} are assumed for ρ and g respectively.

Profile	A (km^2)	p (km)	A/P (m)	$\sin \alpha$	τ_{av} (bars)	Max. depth, Z_m (m)	$(A/P)/Z_m$
98	2.30	5.85	393	0.0240	0.84	535	0.73
101	2.15	5.40	398	0.0265	0.93	575	0.70
104	2.70	5.60	481	0.018	0.77	660	0.73

The mean shear stress for the three profiles is therefore 0.85 bars, which is in agreement with values found elsewhere. This term is a measure of the "average yield stress" for a particular glacier, and from determinations made so far it appears that the average yield stress does not greatly vary from one glacier to another, even under considerably different regimes. The values in the last column are related to the shape of the transverse profiles, and it can be seen that the "shape factor" remains relatively constant for the main portion of the glacier.

The cross-sections, together with movement observations, can also be used to decide whether a particular glacier is advancing, receding or in a steady state. Annual discharge of ice through a transverse section balanced against the total ice loss in the ablation zone should determine the regime of the glacier.

In the case of Gilman Glacier, movement data were obtained over a period of more than a year for some parts of the glacier. Most accurate measurements were obtained near seismic profile 101, where the average yearly maximum rate of movement was 25 m/year near the centre of the glacier decreasing to 21 m/year about 1 km from the margin⁽⁶⁾.

Furthermore, an analysis of the vertical velocity gradient on the basis of plastic flow theories shows that the vertical velocity change is not more than about 2 m/year with most of this change taking place in a narrow zone near the glacier bed⁽⁴⁾. From these considerations a reasonable estimate of the mean rate of discharge through profile 101 is 22 m/year. The cross-sectional area is 2.15 km^2 . The total annual discharge of ice through profile 101 is then $4.7 \times 10^7 \text{ m}^3$. The glacier surface area downstream of this section is 70 km^2 ($= 7.0 \times 10^7 \text{ m}^2$). Thus, a mean annual ablation of 67 cm of ice over the whole of this surface would balance the movement of ice downstream. Ablation measurements in the budget year 1957-58 indicate that the mean ablation over the ablation area was somewhat higher than this value⁽⁸⁾. Since the present movement of the glacier represents certain past conditions of stress, it may be inferred that the present rate of ablation is higher now than in the recent past; this tentative conclusion is supported by the observation that summer temperatures in the North Atlantic area have been generally higher in the last few decades⁽⁹⁾.

3. GRAVITY SURVEY

3.1. Establishment of Gravity Stations

The Dominion Observatory gravity network includes stations at Fort Churchill, Resolute, and Alert. Dr. F. S. Grant, a member of the Operation "Hazen" IGY team, established a series of new absolute gravity stations by tying his network into that

the Dominion Observatory. His stations included one at the base camp on Lake Hazen, one at the Gilman Glacier camp, and one on the east shore of Clements Markham Inlet, which were also stations of the local network. The local stations were chosen at sites where the bedrock elevations had been determined from previous seismic soundings^(3,4). The gravity values were thus all established in absolute terms, although relative changes in gravity, due to the mass deficiency of the ice would have been sufficient for ice thickness computations alone. The gravimeter used was a standard Worden model. Reference was made to previous work along the same lines^(10,11,12,13).

3.2. Reduction of Gravity Values

The ice thickness was computed directly on the assumption that the rock and ice under the gravity station form a slab of infinite extent. It has been shown elsewhere⁽¹⁴⁾ that the thickness (t) can be computed from the equation,

$$t = 1.321 (C_B - 0.1952h),$$

- where
- C_B = Bouguer correction for free air, rock and ice in milligals
 $= \gamma - g_o - A - C_T$,
 - h = surface elevation of the ice in metres,
 - γ = theoretical gravity in milligals
 $= 978,049 (1 + 0.0052884 \sin^2\varnothing - 0.0000059 \sin^2 2\varnothing)$ where \varnothing
 $=$ latitude,
 - g_o = observed gravity in milligals,
 - A = regional Bouguer anomaly in milligals,
 - C_T = terrain correction in milligals.

Densities for rock and ice of 2.71 and 0.9 g cm⁻³ respectively were assumed.

The regional anomaly and the terrain correction are values which, in mountainous regions without exact contour maps, might be extremely difficult to estimate. However, in the case of the Gilman Glacier, because of its width, and because of the uniformity of the surface geology, the regional anomaly can be determined at its centre from a single seismic sounding. It has been shown that, since the marginal nunataks are relatively far away, the terrain correction at the centre of the glacier becomes negligible, and one depth-sounding along the centre line every few kilometres is sufficient to plot the regional anomaly curve as it changes relative to the mountain ranges (Fig. 5). The value of the regional anomaly at the location of interest can then be taken from the curve.

3.3. Comparison Between Seismic and Gravity Profiles

The ice thickness was computed for all the gravity stations and the bedrock elevations were plotted. Comparison between bedrock profiles, obtained by the seismic method, and those computed from the gravity values showed excellent agreement over the central part of the glacier. Even for profiles across the glacier the agreement was good, except at stations near the glacier edge where the discrepancy, due to the uncorrected terrain effects, became large (Fig. 6).

Some of the seismic profiles shot on the ice cap, where firn and ice was some 800 m deep⁽³⁾, gave very poor reflections. The reduction of the gravity observations, taken over the same profiles, revealed a very undulatory bedrock floor, which was not suspected under the level snow surface, but which was apparently the reason for the poor reflections.

3.4. Conclusions

Gravity measurements alone, without seismic depth control, can give a good picture of the shape but no accurate elevations of the bedrock. When supplemented by a few seismic soundings, the gravity method provides an excellent yet simple means of determining the shape and elevation of the bedrock beneath glaciers and ice caps. But, when no contour maps are available and terrain correction is impossible, the easy application of the method to valley glaciers is limited to geologically uniform regions with relatively wide and level glaciers. Although these conditions rarely occur in the glacierized regions of the temperate zones, they are common in the polar regions. Without elaborate terrain corrections, the gravity method fails to give the exact shape of the walls near the edge of a valley glacier.

4. SURVEY CONTROL

Geophysical investigations in areas not yet covered by accurate topographic maps need an adequate programme of survey control. A finer control is required for gravity than for seismic work.

For the reduction of gravity observations in the Gilman Glacier area it was necessary to establish latitudes correct to 0.1 minutes, and elevations between adjacent points correct to 0.3 m. A topographic map showing the relief of the glacier margins was required for estimating terrain corrections.

There was only one previously located point in the area. This was a "Shoran" station on Johns Island, the most northerly station of the Canadian "Shoran" network, some 45 km south-west of Gilman Glacier. The present survey was tied to this station.

The method of extended base, subtense bar traversing seemed best suited for providing survey control for geophysical profiles over wide areas of the ice cap. This method had been used previously on the inland ice of Greenland by Expeditions Polaires Françaises (¹⁵), and by the British North Greenland Expedition (¹⁶). It gives vertical and horizontal control of sufficient accuracy for geophysical work. The rapidity of the method and the lightness of the equipment made it well suited to a small exploratory party. The total weight of surveying equipment was 40 kg, of which no single item weighed more than 8 kg. During 1957 and 1958, a 160-km traverse was made from Chandler Fiord, 28 km south of Lake Hazen, to Clements Markham Inlet, on the Arctic Ocean 60 km north-east of Gilman Glacier; the traverse between the two sea levels closed in height to 1.2 m. Care that reciprocal vertical angles were observed simultaneously contributed to this result.

The survey programme included glacier movement observations, which were based on a local triangulation net extending up the sides of Gilman Glacier. The rate of surface movement is shown in Fig. 1. From the triangulation stations, prominent peaks were fixed, and this control formed the basis for a topographic map. The highest mountain in the area is situated about 20 km west of Mount Oxford and its height was found to be 2500 ± 100 m. It is also believed to be the highest peak in the Queen Elizabeth Islands.

The triangulation net checked a part of the subtense bar traverse that ran from the terminus of Gilman Glacier to a small nunatak 13 km east of Mount Oxford. In a distance of 38 km the horizontal misclosure was 1 part in 3,600, and the vertical misclosure zero. This suggests that the desired accuracy was attained.

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THICKNESS OF THE GLACIERS OF KAZAKH S.S.R. IN ACCORDANCE WITH THE CALCULATION METHODS AND SEISMIC MEASUREMENTS

N.N. PALGOV (U.S.S.R.)

SUMMARY

The paper deals with the calculation of the thickness of a number of glaciers situated in the Zailiysky and Dzhungar Alatau mountains.

Three methods were used, namely those of Lagally, balance and seismic measurements. The method of balance is a new one. It was first published by the author of the paper in 1956 (vestnik of the Kazakh Academy of Sciences. Vol. 4).

The method is based on the definition of the glacier solid flow within any period of time (not less than a month) concerning the profile which is the ground of the greatest profile thickness to be calculated. The required data are as follows: width of the profile 2) shifting of its cross-section for the given period of time (it is defined as the surface velocity multiplied by 0.64 which is the common denominator) 3) lowering of the surface of the glacier on the whole profile (or rising in case of a winter period) expressed by a volumetrical unit 4) Value of the melted mass of the solid ice.

These data are used to define the glacier solid flow (Q) on the given profiles. The latter is represented as the difference between the volume of the melted mass (point 4) and that of the contracted mass (point 3), if no melting occurs then it is equal to the additional ice (point 3 in brackets).

The Q volume and shifting (v.) are taken to define the cross section.

F (profile area) = $Q : 0.64 v$.

The definition of the thickness is done by equating the above given equation to the trapezium or semiellipse.

The remade equation is as follows.

$$H = \frac{14 F}{11 \text{ width of the profile}}$$

where H is the thickness of the glacier, the width of the glacier being the big axis and its thickness is the small one.

The thickness of average glaciers in Kazakhstan of 5-7 kilometres long does not exceed 158 metres.

Comparison done by the three above mentioned methods has been made for certain glaciers.

The results more or less coincide. It should be noted that the comparison of the Lagally and balance methods with that of seismic measurement is not sufficient. However, it is certain that the first two taken together can give quite acceptable results.

As to the method of balance the common denominator (0.64) of the surface shifting of the glacier to the value of the average shifting of the whole cross section for a given profile is not quite defined. This coefficient is not known yet. It is likely to be different for various glacier sections.

In such cases the value of the thickness of the glacier defined by the Lagally equation may be used in the balance method to define the common denominator. Consequently the equation may be represented as follows:

$$K \text{ (coefficient)} = \frac{14 Q}{11 w. H. V.}$$

The paper deals with the concrete data as to the thickness of the glaciers in the Kazakh S. S. R., defined by three above mentioned methods and the results of the comparison.

RÉSUMÉ

La définition de puissance se fit pour une série de glaciers, situés dans les chaînes de montagnes Zailiysky Alatau et Djungarski Alatau.

L'on applique pour cela trois méthodes: celle de Lagally, de bilans et de sondages sismiques.

La méthode de bilans est une méthode nouvelle. Elle fut publiée pour la première fois par son auteur en 1956 (Vestnik Akademii Nauk Kazakhskoi S.S.R., N° 4).

Cette méthode est fondée sur la définition du déversement naturel du glacier pour n'importe quelle période, (mais d'une durée non moins d'un mois) pour l'alignement, sur lequel l'on doit calculer la plus grande puissance de son profil. Comme données nécessaires servent : (1) la largeur de l'alignement; (2) le déplacement du profil de l'alignement en question pour la période indiquée (elle est définie par la vitesse de surface, multipliée par 0.64 — indice généralement admis de réduction); (3) la baisse de la surface du glacier sur toute la parcelle en aval de l'alignement (ou bien la montée, si la période des observations tombait sur la saison d'hiver), exprimée en une mesure volumétrique; (4) la valeur de la masse de glace naturelle fondue.

Le déversement naturel (Q) du glacier sur l'alignement en question représente la différence entre le volume de la masse fondue (art. 4) et le volume de la masse amoindrie (art. 3) ou, en absence de fonte, il est égal au volume de l'apport de la glace sur la même parcelle (art. 3 en parenthèses).

En accord à la valeur Q et au déplacement (V_{moyen}), l'on trouve la troncature du profil.

$$F/\text{aire du profil} = Q : 0.64 V.$$

Ensuite, en égalant la forme du profil à un trapèze ou à une demi-ellipse, l'on calcule la puissance.

Selon la formule convertie de l'aire de la demi-ellipse, dans laquelle la largeur du glacier sert de grande axe et sa puissance — de petite demi-axe, la puissance (H) est définie comme il en suit :

$$H = \frac{14 F}{11 \text{ largeur}}$$

La puissance des glaciers de Kazakhstan (moyens en volume), qui ont une longueur de 5 à 7 km, n'excède point 158 m.

L'on a comparé pour certains glaciers leurs puissances, définies par les trois méthodes, indiquées ci-dessus.

Cette comparaison a montré que les résultats de toutes les trois méthodes coïncident plus ou moins.

Les comparaisons effectuées entre la méthode Lagalli et celle de bilans avec la méthode des sondages seismoque sont encore insuffisante; mais l'on peut affirmer avec assurance que les deux premières d'entre celles-ci en combinaison réciproque peuvent donner des résultats tout-à-fait acceptables.

Dans la méthode des bilans l'indice de réduction 0.64 (du déplacement superficiel du glacier) contre la valeur moyenne du déplacement de tout le profil de l'alignement transversal n'est point fixe assez fermement. Cet indice n'est encore point assez connu, il est possible aussi qu'il est variable pour diverses parcelles du glacier.

Dans ces cas-là, la valeur de la puissance du glacier, trouvée selon la formule de Lagalle, peut être appliquée dans la méthode des bilans pour définir l'indice de réduction. Pour ce but la formule des bilans est convertie en la suivante :

$$K (\text{coefficient}) = \frac{14 Q}{11 \text{ larg. } H \cdot V}$$

L'on donne dans l'article des informations concrètes sur la puissance des glaciers de Kazakhstan, définie selon les trois méthodes indiquées, et l'on cite les résultats des comparaisons.

In the study of glaciers, an important and necessary task is the determination of their thickness.

This task, besides the expensive and complicated methods of boring, seismic soundings and other, is being solved also by means of different calculations, based on these or other indirect data.

One of such calculation methods is the known method Lagally. Its application requires information about the yearly speed of the glacier, the inclination angle of its surface, viscosity and density of the ice.

By some generalization of the data regarding viscosity and density of the ice, the calculation of the thickness by the method of Lagally is simplified and is obtained by the formula:

$$H = 8.4 \sqrt{\frac{V}{\sin \alpha}},$$

where H = the thickness of the glacier in meters, α = the inclination angle of its surface.

In 1956 the author of this article proposed a new method of determination of the thickness of mountainous glaciers that he named the method of balances. This method was published in the magazine "Herald of the Academy of Sciences of the Kazakh S.S.R., no. 4, 1956 and in the magazine" Information of the Allunion Geographical Society, no. 2, 1958.

This method of balances is based on the following considerations.

Moving from above downward the glacier transfers a quantity of substance in the same way as a river does. The volume of the transferred substance (by analogy with the river) represents the natural discharge of the glacier. The word "natural" shows that the discharge concerns solely the ice mass. The transferred substance of the glacier over any profile (the natural discharge) should be shown below this profile by a corresponding increase of the glacier mass. This increase or inclusion can be discovered by means of two levellings of the glacier surface (before the beginning of the observations and after). But, if during the period, for which the natural discharge is being defined, a melting of the surface of the glacier took place, it is then necessary to take into consideration the fall of the surface, caused by the melting.

In the latter case the inclusion (which is practically the natural discharge) will be defined below the given profile, as the difference between the melted (A) and the actually (the discovered by levelling) disappeared volumes of ice (C), that is $A - C = Q$ (the inclusion).

Having defined the value Q on the given profile, then should be found the area of this profile.

For this purpose we consider Q as a product of the profile area and the distance of flow of the glacial mass during the time of observation, which may be of any duration, but not less than one-two months.

The flow distance is defined by observations of the surface speed of the glaciers. It is averaged for the entire profile, and under the consideration of the fact that near to the bottom the speed decreases, is brought to the average of the entire profile by means of multiplication by 0.64 (a more or less commonly-adopted coefficient).

Hence the area of the transverse profile amounts to:

$$F = \frac{Q}{0.64 v}.$$

Adopting the form of a glacial profile similar to a semi-ellipse, we suppose that its large axis is the width (W) of the glacier on the given profile, and the small semi-axis—the maximum thickness (H) of the glacier on the same profile.

From the transformed formula of the area of ellipse, we find that

$$H = \frac{4F}{\pi \cdot W} = \frac{4F}{22/7 \cdot W} = \frac{14F}{11 \cdot W}$$

By methods of balances and formula Lagally there were determined the thicknesses of some glaciers of Kazakhstan, situated in the ranges of the Djungar Alatau and Zailiyski Alatau. These determinations, beginning with the observations and ending with calculations, were carried out by the author, as well as by other explorers:

No. in rotation	Glaciers	Dimensions of glaciers		Transverse Profiles			Thickness on profile, M by methods		
		Length	Area Km ²	No. in rotation	Distances, Km		Balances	Lagally	Seismic soundings
					From end of glacier	From the firm line			
1	2	3	4	5	6	7	8	9	10
1.	Glacier Satpaev (during a year's period)	6,2	10,2	1	1,0	2,0	57	55	—
				2	1,3	1,7	63	57	—
				3	1,6	1,4	78	67	—
				4	3,0	0,0	128	100	—
2.	Glacier Berg (during 375 days)	6,5	14,4	1	1,0	2,3	104	66	—
				2	1,6	1,7	86	81	—
3.	Glacier Satshukin (during 385 days)	4,6	7,2	3	3,3	0,0	76	73	—
4	Glacier Gerasimov (during 392 days)			1	0,7	2,2	67	73	—
5.	Ditto (during 38 days)	4,2	7,1	1	0,5	2,4	158	49	—
				2	0,9	2,0	111	76	—
				1	0,5	2,4	135	49	—
				2	0,9	2,0	117	76	—
6.	Glacier Abaya (during 370 days)	8	12,2	3	1,3	1,6	107		
				1	0,5	4,0	78	62	—
				2	1,0	3,5	99	82	—
7.	Glacier Djambul (during 340 days)	5,5	20,6	3	2,1	2,4	121	109	—
8.	Glacier Shumsky (during 340 days)	3,6	4,1	1	0,6	2,8	91	61	—
				1	0,4	2,0	45	39	39
				2	0,9	1,5	56	68	59

Note 1. — In column no. 2 in brackets is shown the duration of observations used for determination of thickness by method of balances
 Note 2. — In columns 3 and 4 the glacier dimension are given exclusive of dead parts, buried under the frontal moraines.

In the Djungar Alatau the thickness was determined in different parts of the longitudinal profile of eight glaciers (Table 1). On the glacier Satpaev the observations and calculations by methods of balances and Lagally were carried out by N.N. Palgov in 1953, on the glacier Berg by K.G. Makarevich in 1956, and on the remaining six glaciers by P.A. Cherkasov in 1958.

As it is seen from table 1, the methods of balances and Lagally in 9 cases out of 18 give more or less close results with divergencies from each other not more than 20%. Considerable divergencies are noted in the thicknesses of 3 glaciers: Berg, Gerasimov and Djambul, and by all of them, in the lower stream, near the ends.

For the glacier Gerasimov the calculation of the thickness by the method of balances was done for the data of two observation periods (approximately 392 days) and a single month (38 days). The results of both periods differ from each another on the first profile by 14%, on the second—by 5%.

The method Lagally for this glacier gives a particularly large divergence with the method of balances. The calculated thickness by this method is three times smaller on the first profile and 1.5 times on the second profile. For the glaciers Berg and Abaya the thickness in the lower stream is obtained by method Lagally, in comparison with the method of balances, 37-33% less. The authors that carried out the calculations consider that in the lower part of the mentioned three glaciers there is available a damming—up of the stream, caused by a constriction of the bed by the mountain slopes. This damming creates (by a reduced motion speed) an increased thickness of ice. This circumstance, by which a reduction of speed leads to an increase of the thickness, the method Lagally is incapable of considering.

For one of the glaciers of the Djungar Alatau (glacier Shumsky), besides the data of thickness by methods of balances and Lagally, there are also available some data of seismic sounding, carried out by B.A. Borovinsky.

All these three methods gave rather harmonious indexes.

In the ridge Zailiysky Alatau of a great interest are the determinations of thickness of the glacier Zentralny Tuyuksu, where the methods of balances, Lagally, seismic sounding and combined electrical sounding had been applied to (table 2).

The glacier Zentralny Tuyuksu was the object of a complex research by the Geographical Department of the Academy of Sciences of the Kazakh SSR in the period of the I.G.Y. and I.G.C. The observations of the altitude of its surface and the speed of motion were carried out by N.J. Barvenko, observations of the melting of the surface by E.M. Kalmynkina. According to the data of these observations that were taken for the period since 21 July 1957 to 9 January 1958, the thickness of the glacier by method of balances (by appliances of $C = 0.64$) and Lagally calculated by N.N. Palgov.

It ought to be said, that the data concerning the altitudinal position of the glacier surface and its motion, were, in some cases, insufficient and required interpolation.

This might have influenced the precision of the calculation results by both methods.

The seismic sounding for determination of the ice thickness was carried out by the Institute of Physics of the Earth of the Academy of Sciences of U.S.S.R., the combined electric sounding by B.A. Borovinsky (Geographical Department of the Academy of Sciences of the Kazakh SSR).

On the third profile of the ice there was in 1958 bored a hole, that reached the bed at the depth of 52 meters.

The points of determination of thickness on each profile are placed approximately at its middle part, where the largest depth usually occurs.

The comparison of the results, represented in table 2, shows that not all methods appear to be equivalent. With a divergence of 20%, the indexes obtained by method Lagally are in four cases out of six near to the indexes of each of the other methods.

TABLE 2

Thickness of the Glacier of Ridge Zailiysky Alatau

No in rotation	Glaciers	Dimensions of glaciers		Transverse profiles			Thickness on profile (M) by methods				
		length (Km)	Area Km ²	No in rotation	Distance, km		Balance	Lagally	Seismic sounding	Combined electro-sounding	
					From End of glacier	From firm line					
1	2	3	4	5	6	7	8	9	10	11	
1.	Glacier Constitution	4,6	5,9	1	0,1	2,8	—	31	—	—	
10.	Glacier Zentralny Tuyuksu (during 173 days)	3,70	3,5	2	2,7	0,2	—	158	—	—	
				1	0,20	2,28	—	—	32	33	
				2	0,38	2,10	36	38	42	(43)	
				3	0,60	1,88	53	58	52	54	
				4	1,06	1,42	59	69	50	59	
				5	1,62	0,86	84	87	74	83	
				6	1,88	0,60	142	90	100	135	
7	2,48	0,00	119	80	113	115					

Note: The thickness of the Glacier Constitution was determined by G. A. Avsiuk during personal observations in 1942.

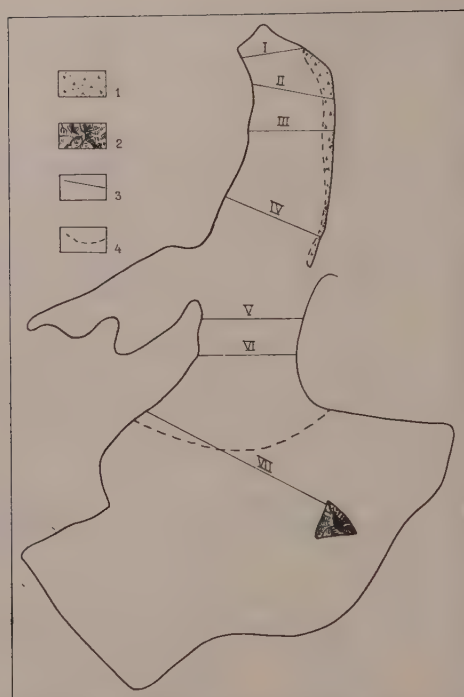


Fig. 1 — Scheme of disposition of profiles on the glacier Zentralny Tuyuksu.
1 — moraines, 2 — Rocks, 3 — Profiles, 4 — Firn line.

and the indexes of the method of balances in five cases out of those six are near to the results of seismic sounding and in all cases to the indexes of electrical soundings. The data of both geophysical methods are more or less equal in six cases out of seven.

The similarity of a number of indexes is clearly shown in the table 3, where a similar indexes have been adopted such that have a divergence of not over 20%.

TABLE 3

Percentage of similar indexes of thickness of glacier Zentralny Tuyuksu by different methods

METHODS				
	Balances	Lagally	Seismic Sounding	Electrical Sounding
Balances	—	67	83	100
Lagally	67	—	67	67
Seismic sounding	83	67	—	86
Electrical sounding	83	67	86	—

The thickness of the glacier on the third profile, ascertained by boring, coincides well with the data of the seismic sounding, the method of balances and electric soundings, and somewhat worse (with a divergence of 10%), with the Lagally results. Hence arises the question—which of the methods appears as most satisfactory? Table 3 shows that the answer to this question is given by the methods of balances and electric sounding.

Not bad results were given also by the seismic sounding. However, by application at the glacier Zentralny Tuyuksu, on the sixth profile it did not reveal the actual character of thickness. Besides the quoted indexes, this is also proved by the geomorphological peculiarities of the glacier bed. Between the sixth and fifth profile the glacier flows in a strongly narrowed bed. Here, on its surface, there is a considerable net of transverse crevasses. All this speaks of the presence on the bottom of the bed of a stony projection—riegel. The existence of the latter was confirmed by the magnetic surveys, carried out by B.A. Borovinsky (the Geographical Department of the Academy of Sciences of the Kazakh S.S.R.). The contoured riegel by the magnetic surveys lies with its comb approximately on the line of the fifth profile, and with its foot on the line of sixth profile. The indexes of the glacier thickness on these profiles obtained by methods of balances and electric sounding, very well reveal this riegel by the corresponding elevations of glacier bed (Table 4).

TABLE 4

Absolute altitudes of glacial bed of glacier Zentralny Tuyuksu on 5 th and 6 th profiles

Profiles	Elevation of glacial bed, (m) by methods		
	Balances	Electric sounding	Seismic sounding
5	3540	3541	3550
6	3525	3532	3567
Altitude of riegel	15	9	Riegel not revealed

The equivalence of the methods of balances and electric soundings is illustrated also by the nearness of elevation results of the bed, as is seen in the table 4.

From this it follows that by studying the thickness of glaciers, the method of balances can fully replace the seismic sounding and electric sounding. As regards the method Lagally, its utilization requires consideration of geomorphological peculiarities of the glacier, particularly cases of the damming up of the stream by narrowing of valley, riegels, etc.

As a control, the Lagally method in these cases can assist the balances method. On its part, the Lagally method can assist a more precise application of the balances method on those parts of the glacier, where it is able to give satisfactory results. The expression "more precise" bears in mind the circumstance that the application in the balances method of the coefficient of conversion of the superficial motion speed of the glacier into average speed of the entire profile cannot always equal to 0.64. Then, the glacier thickness, calculated by the Lagally formula, may serve for the determination of this coefficient. For such cases, out of the equation of balances, the following formula has been derived.

$$C \text{ (coefficient)} = \frac{14 Q}{11 W \cdot H V}, \text{ where } Q = \text{the natural discharge of ice, } W = \text{the profile, } V = \text{the velocity of the profile during the observation period.}$$

SONDAGE SEISMIQUE DU GLACIER FEDTCHENKO OBSERVATIONS GRAVIMETRIQUES SUR LE GLACIER FEDTCHENKO

I.S. BERZON, V.A. PAK, V.N. YAKOVLEV and I.Y. LEONTIEV
URSS

RÉSUMÉ

1. Les explorations sismiques sur le glacier Fedtchenko avaient pour but de préciser l'épaisseur de la glace et l'étude du lit du glacier.

2. Les observations furent exécutées à l'aide d'une station sismique portative à 24 canaux du type CC-24-P. Les caractéristiques de fréquence des canaux sont rectilignes dans la région de 30 à 130 hertz.

3. Une méthode combinée — ondes réfléchies et ondes réfractées — fut appliquée, l'épaisseur de la glace ayant été déterminée à l'aide d'ondes réfléchies, et le lit du glacier, par ondes réfractées.

4. Les travaux accomplis ont démontré que le glacier représente un milieu à trois couches : la glace, la couche intermédiaire et le lit rocheux. Dans la partie moyenne du glacier l'épaisseur de la glace atteint 800 m. Dans la partie de la langue l'épaisseur de la glace diminue promptement jusqu'à zéro. L'épaisseur de la couche intermédiaire est à peu près de 400 m. La vitesse de propagation des ondes sismiques dans cette couche est moindre que la vitesse de propagation des ondes dans la glace.

RÉSUMÉ

1. L'étude du champ gravimétrique sur le glacier Fedtchenko durant l'Année Internationale de Géophysique avait pour but de préciser la possibilité d'appliquer les méthodes gravimétriques à la détermination de l'épaisseur de la glace.

2. Les observations furent exécutées à l'aide d'un gravimètre « Norgard » sur des sections transversales du flux du glacier. La précision des observations était de 0,5 mg.

3. Sous l'influence de l'épaisseur de la glace les anomalies atteignent 15 mg.

4. Une interprétation quantitative des anomalies a été exécutée.

5. On a tiré des conclusions sur la possibilité d'appliquer la gravimétrie à l'étude de l'épaisseur de la glace des glaciers de vallées dans la montagne.

During the International Geophysical Year, in the mountain regions of the USSR and in the Antarctic, Soviet scientific workers carried out extensive glaciological research which also included seismic observations. The latter were aimed at studying the morphology of the glaciers, their depth and structure. Part of this work was seismic research on Fedchenko glacier which was carried out by the glaciological expedition of the Institute of Mathematics named after V.I. Romanovski under the Uzbek Academy of Sciences during the 1958-1959 field seasons. Among those taking part in these observations were scientific workers of the Institute of Mathematics V.A. Pak, V.N. Yakovlev, I.Y. Leontiev, A.K. Kegger and N.S. Usmanova. The group was headed by I.S. Berzon, chief of the Department of Seismic Methods of Prospecting at the Institute of Physics of the USSR Academy of Sciences.

The results of an analysis of the data obtained there are the subject of the given work.

Seismic observations were carried out on a number of sections located at even intervals along almost the entire length of the glacier—in the lower reaches of the glacier (the tongue), in the vicinity of Bivachni glacier, below the high altitude observatory «Lednik Fedchenko» and in the firn zone close to the Tanimasskaya Lag glacier.

The seismic work carried out in 1959 was supplemented with gravimetric observations.



Fig. 1 — Scheme of disposition of observation profiles.

During the entire period of scientific research on the glacier, about 30 km of seismic profiling was done and 118 gravimetric points calculated (Fig. 1).

It has been thanks to the implementation of seismic and gravimetric methods possible to provide a more or less detailed characteristic of the structure of Fedchenko glacier with a detailing of some parts.

1. PHYSICO-GEOLOGIC CONDITIONS OF WORK

Fedchenko glacier is a valley-type mountain glacier which stretches meridionally from the foothills of the Yazgulem ridge almost 70 km in the form of a solid ice stream 2 to 2.5 km wide. Along the entire length, the main ice stream takes in numerous tributaries, which by themselves are sizeable glaciers. Below the firn zone, at approximately in the middle reaches of its stream, Fedchenko glacier receives a big mass of ice from Akademia Nauk, Nalivkin, Elena Rozmirovitch and Kasha-Ayak glaciers.

In its middle reaches, Fedchenko glacier is sharply bent in the form of a horizontal flexure. This sharp bend results in intense cracking of the ice, which is particularly pronounced above the high-altitude observatory «Lednik Fedchenko».

In its lower reaches, Fedchenko glacier receives but little ice from its tributaries.

The passage of the glacier along the slope is hampered by transversal rocky spurs. For instance, in the lower part of the glacier, the influence of the spurs on its morphology is seen very clearly. There, the glacier is extremely narrowed and its width does not exceed 1600 meters. Below this narrow passage, the glacier spreads out fan-like and forms a tridactylous ending which breaks off in steep icy precipices.

Until seismic research was carried out on Fedchenko glacier, its depth was determined indirectly — by measuring the velocity of the ice movement. Such research was realised by R. Finsterwalder in 1928. By using photogrametric methods and by measuring the parallax during repeated observations caused by the movement of the glacier, R. Finsterwalder measured the depth of the ice along a number of profiles. Four kms above Shpora, where maximum velocities of the ice were observed, the explorer, using the Lagally formula, estimated the depth of the ice at 550 meters. In the vicinity of the rocky spurs, near the lower reaches of the glacier, the foot of the ice, according to R. Finsterwalder, was 360 meters deep.

Seismic research carried out in 1958-1959 made it possible to elucidate the question of the depth of the ice in Fedchenko glacier and to answer a number of other questions connected with its morphology and nature of the structure of the bedrock.

2. OBSERVATION METHODS AND INTERPRETATION

2.1. *Seismic exploration*

2.1.1. *Apparatus and observation methods*

Seismic oscillations were recorded by a 24-channel portable seismic station made in the USSR. The total weight of the station including the power supply (alkaline storage batteries), the connecting cable and seismographs is 300 kg. In working conditions, the seismic station was transported either by hand on a sledge. The record of seismic waves was carried out on an open channel (0-0 filtering, cut-off frequency characteristic 30 and 130 cycles per second, without utilising automatic gain control and mixer. The oscillograph of the seismic station is equipped with galvanometres with a natural frequency of 130 cycles per second hertz. The seismographs used in observation were of an electrodynamic type with a natural frequency of 33 cycles per second. They were installed on the surface of the ice in small holes, not deeper than 20 cm, every 25 meters.

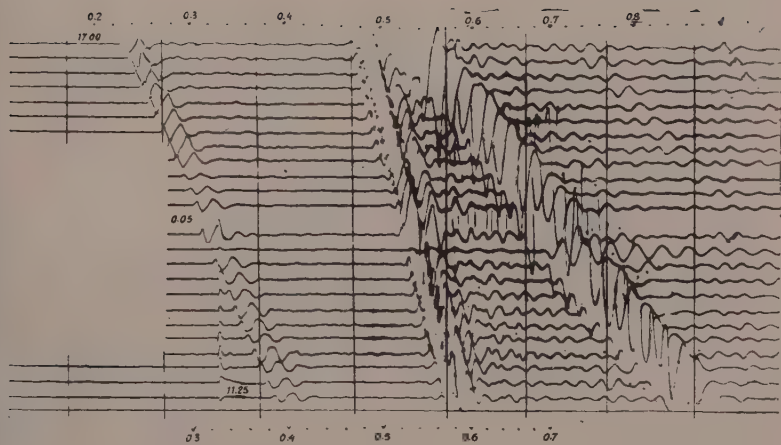


Fig. 2 — Seismogram illustrating recording of waves « t_1 », « t_2 » and « t_4 ».

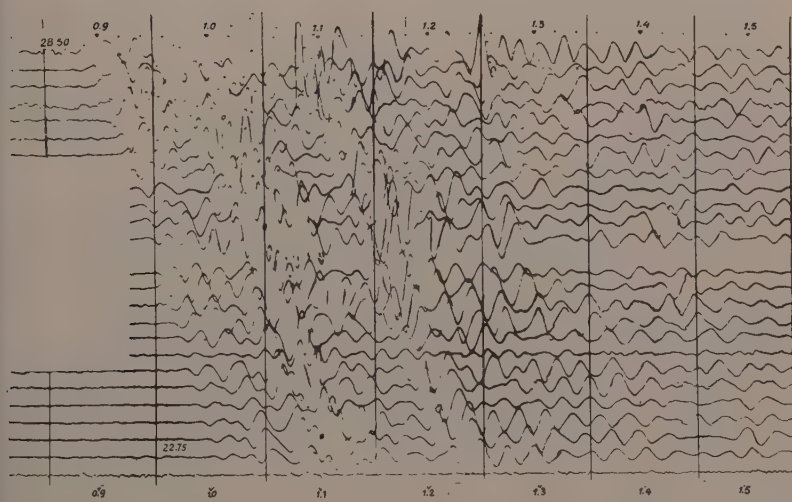


Fig. 3 — Seismogram illustrating recording of waves « t_1 », « t_3 » and « t_6 » at a distance of 4 km from the explosion.

The explosions were held in bore holes, craters and cracks filled with water and directly on the surface of the ice. The size of the charge in most cases was between 0.1-2 kg. In explosions held at some distance from the observation points (3-7 km), the charge amounted to 20-30 kg.

The record of waves reflected from the foot of the ice was done at a close distance from the explosion — not exceeding 1 km. Deflected waves showing the roof of the mother rock at first entrance were recorded at a distance of 3-5 km from the explosion. Observations were carried out along profiles mapped out across the glacier and profiles located along the axis line.

2.1.2. *Methods and interpretation*

The general wave picture is illustrated on Fig. 2-Fig. 5.

The direct lateral wave " t_1 " was recorded in first entrance at a distance of 4 km from the explosion. The loci of this wave are rectilinear and practically parallel. The frequency of the wave " t_1 " was 40-120 cycles per second. The surface Rayleigh wave " t_4 " was recorded only in the following entrances. The intensity of this wave was considerably higher than that of the other waves. The dominating frequency of the wave " t_4 " was 30 cycles per second.

The " t_2 " lateral wave reflected from the boundary ice—intermediate layer could be recorded in following entrances at a distance exceeding the depth of the reflecting boundary four times. The loci of the " t_2 " wave were typically hyperbolic. In transversal profiles, the loci were often distorted with loops arising from the concave form of the glacier bedrock. The frequency of the wave " t_2 " was 50-120 cycles per second.

The deflected wave " t_3 " was connected with the boundary intermediate layer—bedrock. The waves " t_3 " and " t_2 " were sharply distinguished from each other. The wave " t_3 " was of a lower frequency (dominating frequency 30-40 cycles per second) and was characterised by unsteadiness of recording. A change of waves was observed as well as zones of anomalous fading.

The lateral wave " t_5 " reflected from the boundary intermediate layer—bedrock, could not be tracked everywhere. On certain sections (Bivachni and Tanimasskaya Lapa glaciers), the wave was dynamically sufficiently clear-cut. In the middle reaches of the glacier and on its tongue, it could be singled out with great difficulty. Close to the explosion picketage, the wave " t_5 " was not recorded.

Applying effort directed horizontally and across the profile as an experiment, it was possible to record a transversal wave—the direct wave " t_6 " and the wave " t_7 " reflected from the boundary ice—intermediate layer.

In accordance with the recorded loci of the waves, an interpretation was carried out and seismic cross sections were mapped out along two horizons—boundary, dividing ice—intermediate layer (horizon *A*) and surface of the bedrock (horizon *B*).

The loci inclinations of the direct waves " t_1 " and " t_6 " and surface wave " t_4 " have made it possible to determine the speed of the propagation of lateral, transversal and surface waves in the ice close to the surface, which is somewhere in the average of 3700 m/sec, 1800 m/sec and 1700 m/sec correspondingly.

On the basis of the loci of the reflected waves " t_2 ", " t_5 " and " t_7 " and using the method of selection—the theoretical loci, the permanent difference of reciprocal points and effective speeds were determined as well as the seismic horizons "*A*" and "*B*". Along the entire thickness of the ice, the effective speeds, which are 3750-3800 m/sec for lateral waves and 1900 m/sec for transversal waves, differ but little from the velocities determined with the help of loci of direct waves—a fact indicating the homogeneity of the ice. A slight increase in velocities within the ice was recorded in the direction from the upper reaches of the glacier downwards.

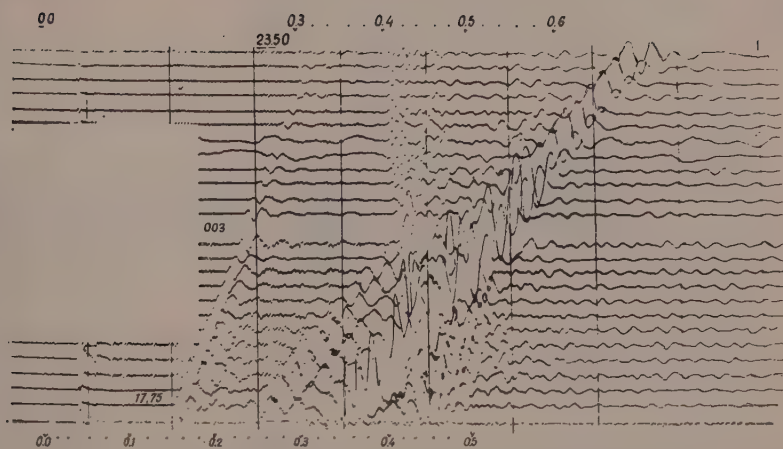


Fig. 4 — Interference of wave « t_2 » at curvilinear boundary of partition (loop effect)

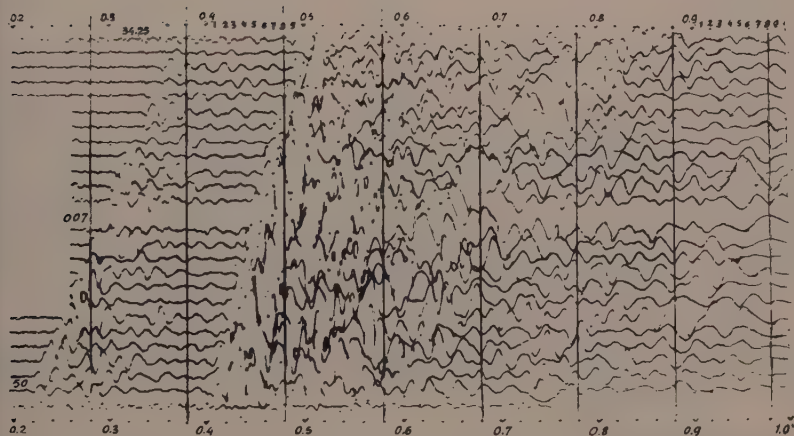


Fig. 5 — Seismogram registering transversal waves « t_6 » and « t_7 ».

The loci of the deflected wave " t_3 " were used for building up the seismic boundary "B" and for determining the velocity of the waves in the bedrock. The velocities along the border of the bedrock were equal to 5500-5700 m/sec.

The loci of the reflected wave " t_5 " have made it possible to evaluate at least approximately the velocity of lateral waves in the intermediate layer. For different sections of this layer, it fluctuates within 3000-3500 m/sec.

2.2. Gravimetry

2.2.1. Apparatus and observation methodes

Research was carried out with a quartz gravimeter. Accuracy of measurements amounted to 0.5 mgl. In order to ensure reliable control of the nil point, the method of 100% repeat was applied. Observation was carried out along profiles transversal to the glacier stream. The distance between observation points in one row did not exceed 200-300 meters.

2.2.2. Density of the rock

The density of the ice was found to be 0.9 g/cm³. The average density of the bedrock (determined on samples) is equal to 2.75 g/cm³.

2.2.3. Correction

All observations have been reduced to the surface of the glacier. The topographic correction, which in conditions of Fedchenko glacier amounts to 5-20 mgl, was calculated with the help of graphs elaborated by P.I. Lukavchenko.

2.2.4. Interpretation methods

Intense negative anomalies, reaching 30-40 mgl were observed on transversal profiles. It was only in the lower reaches of the glacier (tongue), where the depth of the ice is not great, that the negative anomalies did not exceed 15 mgl. The regional background in view of the insignificant length of the profiles, was not taken into consideration.

Quantitative interpretation was carried out by the method of successive approximations.

Using the curve V_z as a basis and the formula $h = \frac{\Delta g}{2\pi f \sigma}$, where f = permanentn gravitational, g = anomaly of force of gravity and σ = excessive density, the depth of the foot of the ice was determined in initial approximation (S_1). The curve S_1 and the G.A. Gamburstev graph make it possible to find the theoretical value of $V_z^{(1)}$ in points of observation. The difference $V_z - V_z^{(1)}$ was summed up with the recorded values of the force of gravity and the new data served as the basis for compiling a profile of the second approximation (S_2). This process was carried on until the difference $V_z - V_z^{(n)}$ became less than the minimum accuracy of observation. This finished the work of making up sections along profiles.

Apart from this, the method of selection was also used.

In the lower reaches of the glacier, where the depth of the intermediate layer is something like 500-700 meters and the thickness of the ice is only 150 meters, the density of the intermediate layer was measured and found to be 2.3 g/cm³.

3. RESULTS OF RESEARCH

The complex utilisation of seismo-gravimetric methods of research have made it possible to provide the following picture of the structure of Fedchenko glacier.

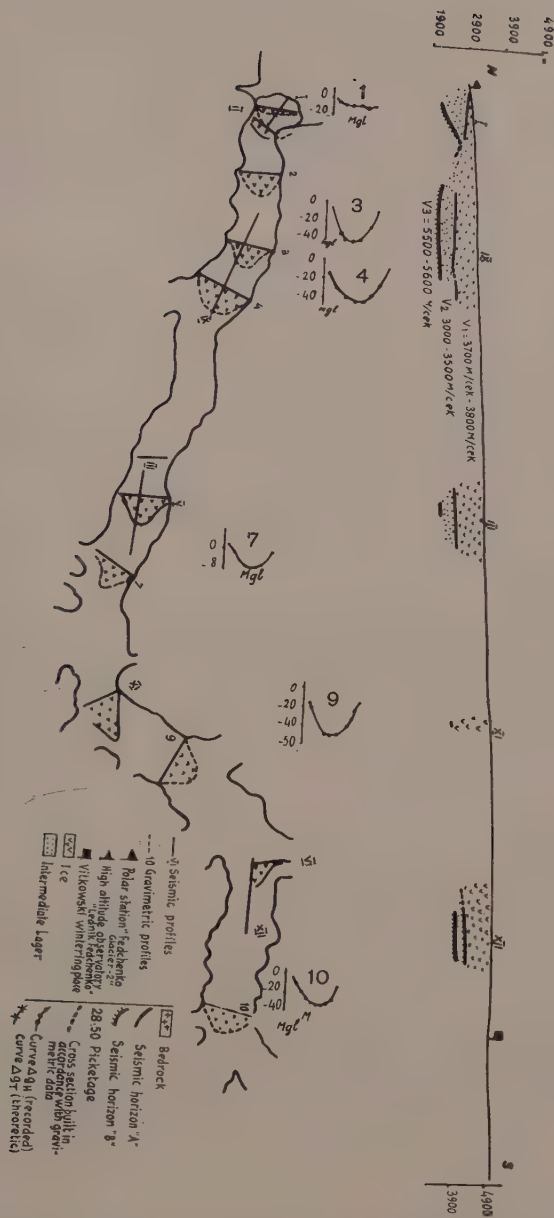


Fig. 6 — Scheme of structure of Fedchenko glacier.

The body of the glacier rests on a thick glacial bed which is the intermediate layer between the ice and the bedrock. The presence of the intermediate layer is recorded wherever observations were carried out by the correlations method of deflected waves (CMDW). The body of the glacier constitutes a homogenous mass of ice characterised by the velocities of lateral seismic waves of 3700-3800 m/sec. The density of the ice is 0.9 g/cm³.

According to CMDW observations, the boundary velocities in the bedrock are 5500-5700 m/sec. Density, estimated as a result of the study of 180 samples is 2.75 g/cm³. Geologic surveying has shown that the bedrock of Fedchenko glacier comprises metamorphosed shale with interlayers of limestone, broken up in places by granite and Paleozoic granodiorites.

In the intermediate layer, the velocity of the lateral seismic waves is less than in the ice and that is why no head waves were recorded from the boundary ice-intermediate layer. The loci of reflected waves connected with the surface of the bed helped to evaluate the velocity of lateral seismic waves in the intermediate layer, which is 3000-3500 m/sec. According to gravimetric data, the density of the intermediate layer is something like 2.3 g/cm³. The above-mentioned data testifies to the coarse-grained composition of material in the intermediate layer.

A general outline of the nature of the glacier's structure is shown on Fig. 6.

The maximum depth of the ice was recorded in the middle reaches of the glacier (1000 meters). On the glacier tongue, the thickness of the ice is only 200-250 meters and further on the ice thins out. Apparently, the glacier has a similar structure in its upper reaches. It is necessary to draw attention to the corrugated nature of the lower boundary of the ice (in lateral direction), which was noted in the vicinity of Bivachni and Tanimasskaya Lapa glaciers.

The intermediate layer is quite evenly spread along almost the entire surface of the bedrock and is something like 300-400 metres thick, with the exception of sections with transversal elevations where the thickness of the layer is sharply reduced.

The cross section of the glacier has been studied with the help of profile shooting. Two types of cross sections have been noted. For sections with thick ice, a general V-shaped form of the glacier bed with a few distorted slopes is very characteristic. The latter creates the impression that there are two trough valleys placed in one another.

The other type are the cross sections of the glacier at its tongue. There, the valley is of a U-shaped form. Apparently, such a form is characteristic of all sections with insignificant thicknesses of ice.

The glacier bed is very assymetric.

A study of the cross sections of the glacier reveals the fact that along its entire length, they narrow and widen alternately, which means that the glacier body has a "bead"-shaped structure. Probably, these peculiarities in the structure of the glacier arise from the maximum cutting of the glacier valley and are the result of the irregular movement of the ice stream.

Knowing the velocity of the propagation in ice of direct, lateral, transversal and surface waves, it was possible to find the value of the Poisson ratio σ for ice which is equal to 0.36. The value of the transverse thrust modulo has also been calculated: $M = 27.3-29.2$ m/sec² as well as the Young modulo: $E = 75.2-78.8$ m/sec².

4. CONCLUSION

1. A cross section of Fedchenko glacier presents a two-layer medium comprising a layer of ice with an intermediate layer beneath it, both resting on bedrock. The thickness of the ice in the middle reaches of the glacier is 700-1000 meters. In the lower

reaches of the glacier, the ice layer becomes less thicker (400-300 meters) and finally thins out completely.

2. The velocity of the propagation of lateral seismic waves in the intermediate layer (3000-3500 m/sec) and the density of the latter (2.3 g/cm^3) testify to the coarse-grained composition of the intermediate layer. Apparently it is a bottom moraine, but it may also be assumed that it is of an alluvial nature.

3. The research carried out on Fedchenko glacier makes it possible to evaluate the effectiveness of studying valley-type glaciers by a complex of geophysical exploration methods. It is recommended to use jointly the method of reflected and deflected waves supplemented with gravimetric surveying.

THE BOREHOLE EXPERIMENT ON BLUE GLACIER, WASHINGTON

Ronald L. SHREVE

University of California, Los Angeles, California, U. S. A.

INTRODUCTION

Blue Glacier is a temperate valley glacier located on Mount Olympus (latitude 47°48' N, longitude 123°42' W, elevation of summit 2400 m. above sea level) in Olympic National Park, Washington. It is fed by a steeply inclined ice fall that descends 300 m. from an accumulation area near the summit of the mountain. The base of the ice fall lies at an elevation of 1800 m. From it the glacier flows 3 km. to the terminus at 1300 m. The width of the glacier decreases from 900 m. near the ice fall to 500 m. at the terminus. In late August the firn limit lies at an elevation of about 1600 m. approximately 1 km. downglacier from the base of the ice fall.

The borehole experiment on Blue Glacier follows the well-established technique of drilling several deep holes vertically into the glacier, simultaneously emplacing aluminium pipes approximately 5 cm. in diameter, and each year making inclinometer surveys of the pipes to measure the deformation they have undergone. Assuming that this deformation is the same as that of the ice surrounding the pipes, and assuming further that the stress field in the ice is the same as that in a glacier of uniform surface slope and depth, and infinite length and width, flowing parallel to its bed, it is possible to deduce the relationship between shear stress and strain rate for comparison with theory and experiment.

The borehole experiment was begun in 1957 and continued in 1958 and 1959 as part of the Blue Glacier research program of the California Institute of Technology, performed in conjunction with the International Geophysical Year, and supported by the U. S. National Committee for the IGY. R. P. Sharp, director of the Blue Glacier project, made and reduced the inclinometer surveys reported in this summary, and all of the members of the project shared the tedious job of drilling the boreholes.

EQUIPMENT AND TECHNIQUES

The boreholes were drilled with electrically-heated thermal boring devices, called «hotpoints», that were specially developed as part of the project. These hotpoints are 5 cm. in diameter and 20 cm. long. They are designed to be operated immersed in water at depths as great as 600 m., but they have never actually been used at depths greater than about 150 m. They dissipate 2400 watts of power at 230 volts. When operated at this wattage, they drill through solid temperate ice at approximately 10 m. per hour, making a hole about 6 cm. in diameter. At this power input these hotpoints will operate thousands of hours without failing.

Power was supplied by a commercial portable 2500-watt 230-volt 60-cycle AC generator driven by a two-cycle gasoline engine that requires about 4 liters of fuel per hour when operated under full load. The entire powerplant weighs approximately 55 kg.

The technique of drilling was to attach the hotpoint to the lower end of the pipe to be emplaced. The power line passed down the inside of the pipe to the hotpoint; and the pipe served as the ground return. The pipe was divided into sections each 3 m. long. A new section was screwed onto the top of the pipe in the hole every 10 minutes on the average. When the hole was completed, the power wire, which

was made to break at the lower end, was removed, and the inclination of the emplaced pipe measured at 8 m. intervals. The inclinometer used was a single-shot Parsons model that, after a pre-set interval of time, records photographically the positions of a compass needle and a pendulum within the instrument. This instrument is capable of measuring inclinations up to 4 degrees from the vertical with an accuracy of 5 seconds of arc. With shorter pendulums it can measure inclinations up to 27 degrees with the same relative accuracy. The inclinometer surveys are repeated each summer for as many years as possible.

A major source of difficulty in subsequent surveys of the Blue Glacier boreholes has been ice that forms inside the pipes despite all preventive measures. Special hotpoints only 3 cm. in diameter are now being constructed to melt out this ice in future surveys.

OBSERVATIONS AND CONCLUSIONS

Five boreholes have been drilled in the course of the Blue Glacier project : two in 1957, two in 1958, and one in 1959. The 1959 hole is located near the valley wall; it has not yet supplied any useful data. The two 1958 holes are located near the centerline of the glacier. They lie 200 m. apart on approximately the same flowline at about 1600 m. elevation, that is, near the August firn limit. The upstream hole was about 80 m. deep and the downstream one 120 m. deep in 1958. The two 1957 holes are located about 60 m. downstream and 75 m. to the right (looking downstream) of the two 1958 holes, respectively. The upstream hole was 225 m. deep and the downstream one 110 m. deep in 1957. The depth of the ice in the vicinity of these holes has been found by the seismic reflection method to be roughly 300 m.

The slope of the ice surface at the 1957 and 1958 boreholes is about 4.5 degrees. The average surface velocity is 13.5 cm. per day. The measured longitudinal strain rate is negligible; however, scattered transverse crevasses exist in the immediate vicinity of the boreholes, and the edge of a large field of transverse crevasses lay only 50 m. downglacier from the lower two boreholes in 1958.

Unfortunately, it has been possible to survey only two of these four holes in successive years to sufficient depth to permit analysis of the flow of the ice.

The lower 1958 hole was surveyed to 120 m. in 1958, and to 110 m. in 1959. Making the analysis described in the introduction (assuming a uniform density of 0.91 gm. per cm³ for the ice), the data for this period, 1958-1959, fit the power law relationship

$$\dot{\gamma} = (\tau/B)^n,$$

in which $n = 2.6$, $B = 2.6$, the shear strain rate $\dot{\gamma}$ is given in years⁻¹, and the shear stress τ is given in bars (1 bar = 10⁹ dynes per cm² approximately). The curve fits quite closely for shear stresses ranging from 0.5 to 0.8 bars, corresponding to depths of 70 m. to 100 m.; for smaller stresses, corresponding to smaller depths, the scatter increases considerably.

The upper 1957 hole was surveyed to 225 m. in 1957, to 150 m. in 1958, and to 100 m. in 1959. The data for 1957-1958 fit the power-law relationship with exactly the same constants, $wf = 2.6$ and $B = 2.6$, as deduced for the lower 1958 hole. The data from this hole for 1958-1959, however, do not give a smooth curve. In fact, they suggest that some parts of the pipe moved upglacier and other parts moved downglacier in an unsystematic fashion. It is hoped that the 1960 inclinometer survey of this borehole will elucidate these peculiar data.

EXPERIENCES WITH ELECTRO-THERMAL ICE DRILLS ON AUSTERDALSBRE 1956-59

W.H. WARD (Great Britain)
(Cambridge Austerdalsbre Expedition)

SUMMARY

An account is given of the design, construction and performance of electro-thermal drills for sinking two lined and two unlined boreholes into the dense ice of a temperate glacier (Austerdalsbre, Norway). The drill heads were 5 in. (12.7 cm) and 3.2 in. (8.1 cm) in diameter and they melted holes only slightly larger than these diameters quite efficiently and reliably. The electrical elements were placed close to the leading face of the drill in an aluminium-alloy casting and they were protected from fusing during the penetration of zones resistant to melting by means of a thermostat. Thermal drilling is seriously impeded by a thin layer of rock debris which accumulates beneath the drill in dirty ice.

RÉSUMÉ

Des indications sont données sur la constitution, la construction et les résultats d'une sonde glaciologique électrothermale pour la réalisation de deux sondages tubés et de deux sondages non tubés dans la glace dense d'un glacier tempéré (Austerdalsbre en Norvège). Les têtes de sonde avaient 5 pouces (12.7 cm) et 3.2 pouces (8.1 cm) de diamètre et elles fondaient des ouvertures à peine plus larges que ces diamètres avec beaucoup d'efficacité et d'aisance. Les éléments électriques étaient placés tout près de la face guidante de la sonde dans un dispositif en alliage d'aluminium et ils étaient protégés contre la fusion pendant la pénétration de la sonde dans des zones résistantes à la fonte à l'aide d'un thermostat. Le sondage thermique est sérieusement gêné par une mince couche de débris rocheux qui s'accumulent sous la sonde dans la glace sale.

In 1955 it was planned to lay a pipe horizontally along the tunnel (Glen, 1956), which had been dug in the Odinsbre ice fall (Austerdalsbre, Sogn, Norway), in order to continue measuring the movements of its axis after closure of the tunnel. An aluminium-alloy pipe with an internal diameter as large as 3 in. (7.6 cm) was selected to allow reasonably accurate surveys to be made of the almost horizontal pipe axis with a special pendulum inclinometer. Unfortunately this expensive pipe, 420 ft (128 m) long, was not delivered in Norway in time for laying in the tunnel, so in subsequent years it was sunk into the glacier by means of electro-thermal drilling. The diameter was unnecessarily large and the pipe very cumbersome for this purpose, but the large diameter did facilitate the development of a reliable and efficient type of electro-thermal drill.

This paper describes the design and construction of the drill and its performance during the sinking of two lengths of pipe into dense glacier ice, one in 1956 and the other in 1958; the electro-thermal boring of two unlined holes in 1959 with a drill of similar design, but of smaller diameter, is also described.

1. INITIAL DESIGN OF A 5 IN. (12.7 cm) DIAMETER DRILL HEAD

The pipe with an internal diameter of 3 in. (11.8 cm) had walls 1/4 in. (6.4 mm) thick and was fitted with screwed-socket joints at intervals of 15 ft (4.6 m). The male and female threaded portions of the joints were welded on to the outside of the tubes so that they were not weakened by screw-cutting, see Fig. 1. The joints were kindly designed and fabricated by the A.P.V. Co. Ltd. To provide some clearance over the

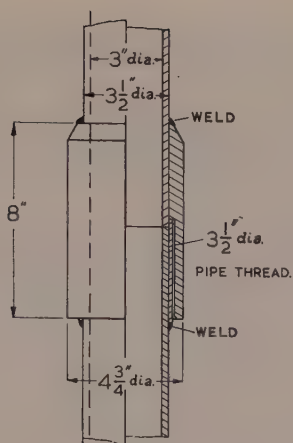


Fig. 1 — Details of joint in 3 inch (7.6 cm) internal diameter aluminium-alloy pipe.

sockets (4.75 in., 12.1 cm dia.) the thermal drill head was made 5 in. (12.7 cm) diameter.

The design of the first drill head for use in temperate ice was completed during the winter 1955-56 with the following principal objects in mind:—

1) the electrical element should be distributed only over the leading face of the drill so that heat is dissipated only directly downwards to melt a hole of the size required (5 in. dia.) and no greater;

2) the drill head is separated everywhere from the ice by a thin rapidly-flowing layer of melt-water, and this water should yield all its available heat to the ice before leaving the leading face of the drill.

In view of 2) above it was decided that the melt-water should flow radially outwards over the leading face rather than drain directly upwards through the heater. In the former case, the warm melt-water would have a greater opportunity for melting ice below the drill before it was released upwards at the periphery of the face. To obtain the greatest heat transfer the thickness of the melt-water layer should be kept thin by the weight of the pipe and the radial flow path across the interface made as long as possible. A simple form of thermal drill illustrating these features is shown in Fig. 2

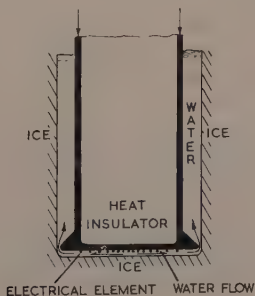


Fig. 2 — Diagram of simple form of electro-thermal drill.

Many practical limitations enter into the design of the electrical resistance element and tend to dominate the whole design of the drill. Initially a simple spiral of nickel-chrome wire mounted flat at the end of a rod and placed directly in contact with the ice was examined(*). Experiments showed that the wire was difficult to support, unless it was quite thick. It also needed a low voltage, a high current, very heavy cables to reduce lead losses and a heavy generator or a transformer. Quite small air or steam bubbles generated in the melting process caused local hot spots and easily led to fusing of the element.

A survey of other materials of significantly greater specific resistance indicated that they would be unreliable on account of oxidation. In these circumstances the only solution was to find a way of winding a considerable length of nickel-chrome wire in a compressed form inside a metal body of high thermal capacity and conductivity close to the leading face of the drill, so as to use about 200 volts, light cables and a light generator. One commercial form of element appeared to be most suitable, namely the type consisting of a helix of nickel-chrome wire embedded in magnesite and sheathed in a solid-drawn metal tube, with the tube itself wound into a tight spiral and cast into a copper or aluminium body.

Some very helpful discussions were held with Mr. D.L. Lewis and Mr. R.J. Locock of the General Electric Co. Ltd. (England) on the design of these elements and they kindly agreed to make several. It was apparently difficult to cast the elements into copper, and as the copper oxide film, which develops rapidly in use, is a poor thermal conductor, we decided to use an aluminium alloy casting. Dr. Richard Seligman of the A.P.V. Co. Ltd., kindly agreed to make these castings.

The first thermal drill which resulted from the co-operation of the above firms and was used in the summer of 1956, is shown in Fig. 3. The leading face was not flat, but was formed into a 90° cone because, i) it permitted a greater length of electrical element to be used and kept closer to the centre of the drill head, ii) it offered better steering qualities, and iii) the warm melt-water path was greater. A path across the

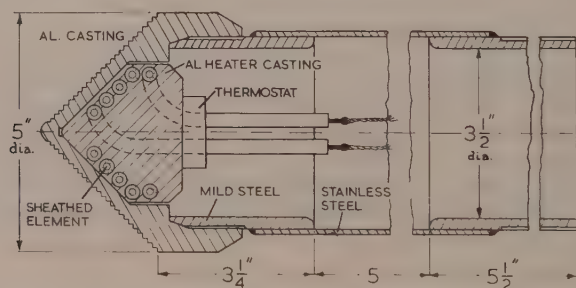


Fig. 3 — Details of the construction of the 5 inch (12.7 cm) diameter drill used in 1956.

leading face was further increased by means of a series of circular grooves of triangular section turned on the conical face. The apparent ease with which many earlier forms of thermal drills had fused suggested that special arrangements were desirable to reduce the risk of fusing, and to make the element replaceable without the need of lifting the pipe out of the glacier. The sheathed element was therefore cast into a separate aluminium cone which fitted accurately inside the outer nose cone, and a

(*) This arrangement was a development of a drill used by Dr. C.A. Bunton in North-East Land in 1955.

thermostat was fitted on the topside of the heater casting. This arrangement also allowed the heater and the cable to be extracted when the pipe had been sunk to its final position.

An annular air gap was formed between the cylindrical side of the heater casting and the outer nose cone to reduce the radial heat flow, and a thin tube of stainless steel was used to join the drill head to the pipe and reduce the heat conduction upwards. Low heat losses from the topside of the heater casting depended on the pipe being kept free of water and special attention was given to sealing the pipe joints on this account.

Three thermal drill heads of identical form, and with element loadings of 1.5, 2.0 and 2.5 kW at 230 volts were made with the object of finding by experiment the maximum loading which could be used safely without the element fusing during ice drilling. These elements were sheathed in a nickel-chrome-iron alloy («Inconel») tube and the effectively-heated length of sheath was about 28 in. (71 cm) in each case. The 2.5 kW element was therefore loaded to about 90 watts/in. (35 watts/cm) of sheath. All three drill heads survived laboratory trials made first in well-stirred ice-laden water, and secondly by melting holes in dense ice; if more time had been available more heavily-loaded elements ought to have been tried. The 2.5 kW drill head penetrated the laboratory ice at a rate of 6.0 ft/hr (1.8 m/hr); the periphery of the drill head remained quite cold and the diameter of the melted hole was not more than 1/8 in. (3 mm) greater than the drill head diameter. If 100% penetration efficiency is defined as the rate of penetration when all the available electrical energy is consumed in melting a hole in the ice of a diameter equal to the drill head diameter, then the laboratory efficiency was about 93%—a very satisfactory result.

2. THE 1956 PIPE PROJECT

Because the tunnel did not reach bedrock at the foot of the Odinsbre icefall a pipe was sunk for this purpose in 1956 at the 1955 site of the tunnel entrance at a slope of about 24° to the vertical and normal to the ice surface. The first pipe length with the thermal drill attached was set and maintained in the correct direction with a guyed tripod, and as more pipe lengths were added this direction was maintained quite well without any special attention.

The electrical generator consisted of a «B. S. A.» 420 cm³, single cylinder, 4-stroke petrol engine weighing 96 lb (43 kg) coupled-in-line by a flexible coupling to a 230 volt, 3 KVA, 50 cycle, self-exciting alternator weighing 110 lb (50 kg), and running at 3000 revs. per minute. The engine and the alternator were separated to facilitate transport from the road head and were mounted together on the icefall on a baseplate built from aluminium-alloy channel sections. This arrangement gave rise to some vibration troubles. The engine regularly consumed 4.0 pints (2.3 litres) of petrol per hour and was kindly loaned by the makers.

The alternator output was fed through a distribution box fitted with switches, fuses, voltmeter and ammeter, and led to the thermal drill by a two-core cable. The cable had flexible cores containing 110 tinned-copper wires 0.0076 in. (0.19 mm) diameter, insulated with vulcanised rubber, and sheathed in tough rubber. It weighed 35 lb. per 100 yds (17 kg per 100 m) and had a «go and return» resistance of about 1 ohm per 100 m of cable.

Observations were made at frequent intervals of penetration, current, voltage and petrol consumption during the pipe-sinking operation. A record of the rate of penetration and the cumulative heat dissipation is given in Fig. 4. For the first 15 ft the pipe entered the ice at the laboratory test rate of 6 ft/hr (1.8 m/hr), but unfortunately the pipe subsequently filled slowly with water, despite the careful attention that was given to the pipe jointing. The water boiled continuously behind the heater, the pene-

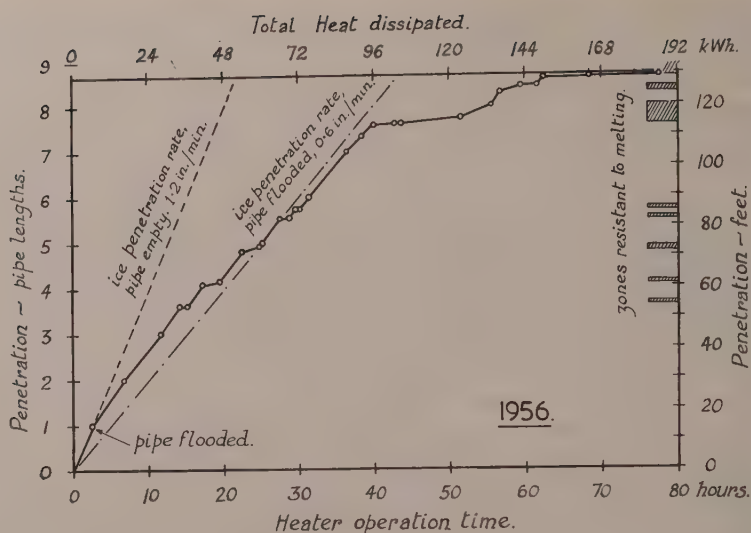


Fig. 4 — Penetration performance during the 1956 project.

tration rate was halved, and the ice hole became enlarged in diameter. There was never any tendency for the ice to grip the pipe sockets despite the rapid movement of the icefall (about 300 m/yr), and the highly compressive longitudinal surface strain-rate of about 1.0 yr^{-1} (Nye, 1959).

The pipe passed through several resistant zones which are believed to consist of ice laden with rock debris. The pipe would suddenly stop sinking, but gentle surging of the pipe up and down, or rotation of the pipe, would cause slow sinkage. Equally, suddenly the pipe would accelerate and penetrate a few inches at its normal speed and then slow down again. It is thought that the heater is prevented from melting ice when quite a thin layer of fine dirt accumulates beneath the drill and that surging of the pipe flushes the dirt away. This process has been demonstrated by melting a shallow hole in dirty ice. Fig. 4 shows the zones which were resistant to penetration and it is thought that a record of this nature is of some value in revealing dirt zones in a glacier. It is, of course, possible that some of the dirt will have come from a higher or thicker zone than is indicated. The operation was finally stopped when no effective penetration was obtained after dissipating 2.4 kW at the drill for 14 hours, and it was presumed that bedrock had been reached at a depth of 129 ft (39.3 m). The hole around the pipe remained full of water throughout the sinking operation; no large voids or crevasses were detected, and whenever the pipe was sinking a steady stream of air bubbles rose to the surface of the water; some of the bubbles were very large.

Overnight when the drill did not operate it was noted on several occasions that the pipe had apparently sunk an inch or so relative to the ice surface, in addition to some small ablation which may have occurred. When the resistant bottom had been reached and the pipe was still quite free in the ice, measurements showed that the apparent penetration was practically the same whether heat was being dissipated or not and it was clear that we were measuring the average rate of strain of the ice normal to its surface. Three measurements taken over periods of about half a day give a mean strain-rate of 0.8 year^{-1} . Although this rate compares well with the value of 0.75 year^{-1} calculated from measurement of the transverse and longitudinal strain rates on the ice

surface (Nye, 1959), it does not necessarily suggest that the pipe reached bedrock since it is not known how the strains vary with depth.

It would have been possible, had we thought of it, to measure the variation in the strain-rate normal to the ice surface by stopping the pipe at various depths. In general it will be necessary to allow for the penetration caused by the additional stress of the pipe (or water) at the bottom of the hole; in our case this correction was negligibly small.

The pipe suffered the same fate as the tunnel entrance did a year earlier. Neither of these prominent features have been seen again (summer 1959). Their disappearance remained a mystery until surface velocity vectors demonstrated that the ice in this area becomes temporarily buried beneath the thick ice-avalanche deposit falling from the adjacent icefall, Thorsbre (Nye, 1959).

3. THE 1958 PIPE PROJECT

The thickness of ice in the Odinsbre icefall proved to be much less than we had anticipated and a considerable length of pipe was left over. In 1958 a further length was purchased to give a total length of almost 400 ft (122 m). This pipe was sunk in the main glacier (Austerdalsbre) approximately 1 mile (1.6 km) below the icefall (*) and in the centre of the valley at a point on a transverse section where the forward velocity is a maximum (about 150 ft/yr, 47 m/yr) and where the longitudinal strain-rate is slightly compressive (about 0.01 year^{-1}). The average slope of the glacier surface in the direction of surface flow and over a distance of the order of the glacier thickness is $4^{\circ}46'$. The pipe was sunk approximately normal to the surface for the purpose of measuring the vertical distribution of velocity parallel to the surface. A gravity survey (Bull and Hardy, 1956) had suggested that the ice thickness at the site was 300 ft (90 m) ± 50 per cent.

Since it had proved impracticable in 1956 to prevent water entering the pipe it was obviously preferable to form a sealed and insulated cavity behind the heater, and allow the pipe to be flooded without reducing the penetration efficiency, but this required some provision for separating the supply cable from the drill when the boring is complete.

The final form of the 5 in. (12.7 m) diameter thermal drill is detailed in Fig. 5. This arrangement was completely satisfactory and the writer would use this design again without any basic modifications for drilling a hole about 5 in. diameter or larger. The heater loading could be increased to some extent without reducing the efficiency.

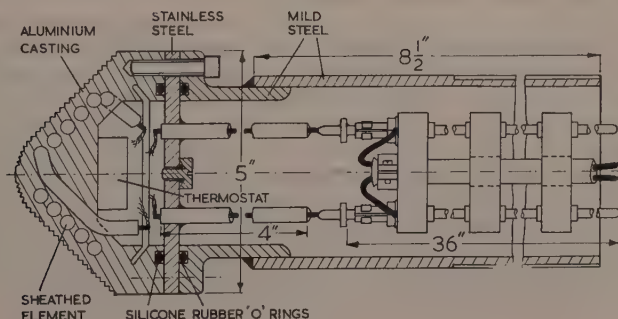


Fig. 5 — Details of the construction of the 5 inch (12.7 cm) diameter drill used in 1958.

(*) For a recent map of the glacier, see King (1959).

The sheathed element is cast directly into the aluminium-alloy nose cone and the thermostat is mounted directly behind it. The flanged coupling is necessary for assembly and the air-tight compartment behind the heater is closed by a stainless steel disc, which is inserted between the coupling, and sealed on both sides by two silicone 'O' rings. The compartment is filled loosely with glass wool, and its airtightness is tested through the screwed plug. The plug is sealed with an 'O' ring and it enables a screw-driver to be inserted for adjusting the thermostat.

The electrical supply is taken through the stainless steel disc by two nickel rods insulated in magnesite and sheathed with «Inconel» tubes, just like the resistance element. The upper end of the magnesite is impregnated with a silicone resin and the recess in the upper ends of the tubes filled with an «Araldite» resin, which expands on setting. The electrical connections inside the compartment are made with flexible nickel cords sheathed with ceramic beads.

The connections to the external cable, which have to be separated when the drilling is complete, is made by means of a special «plug and socket» with a travel of 3 ft, which develops an increasing mechanical resistance to separation in two stages to avoid accidental uncoupling. The «plug» is formed by two brass rods, 3 ft (1 m) long, which are brazed on to the upper ends of the nickel rods. Three «sockets» slide along the «plug», and the cable which passes freely through holes in the centre of the upper two sockets, is anchored and connected to the lower socket. The lower socket holds two split brass sleeves, which grip the brass rods with the aid of spring-steel «C» clips. This method of uncoupling the cable worked entirely satisfactorily.

The element resistance was 16.3 ohms under normal operating conditions and the effectively-heated length of its sheath was about 32 in. (81 m). A brief laboratory trial with the drill loaded to 2.4 kW showed that its ice penetration efficiency was about 80-85 per cent.

The generator used on the glacier in 1958 and subsequently, was similar to the one mentioned previously, but the alternator and engine bodies, and their shafts, were coupled directly. It was mounted in a light tubular framework and was supplied by the Welland Engineering Co. Ltd. The generator with sufficient fuel for the 1958 project and for the subsequent projects in 1959, together with nearly all the 1958 expedition supplies and personnel were lifted by helicopter from the roadhead to the glacier by the Royal Norwegian Air Force.

All the pipe available was sunk to a depth of 397 ft (121 m) in the glacier without any serious problems and without reaching bedrock. The penetration and petrol consumption are plotted against operating time in Fig. 6. It will be noticed that the rate of penetration for the first 200 ft (mean value about 5.9 ft/hr, 1.8 m/hr) is less than for the second 200 ft (mean value about 6.6 ft/hr, 2.0 m/hr). The reason for the change in the mean rate of penetration is not clear.

Throughout the operation the generator voltage was kept steady at 238 volts by means of a rheostat in the field winding and down to a depth of about 240 ft (73 m), when one length of supply cable was in circuit, the generator current remained at about 13.0 amps. During this period it is estimated that the output at the drill was about 2.7 kW and the mean penetration efficiency about 75 per cent. Below this depth, when two lengths of supply cable were in circuit, the generator current was about 12.5 amps, the drill output about 2.5 kW and the mean penetration efficiency about 85 per cent. The change in penetration efficiency does not appear to be associated with some regular change in the melting properties of the ice, since a hole was drilled at the same place (but in different ice) the following year with different results.

The thermostat cut the circuit on several occasions while penetrating a few zones which were difficult to melt; the positions of these zones are shown in Fig. 6. Melt-water overflowed at the surface outside the pipe throughout the sinking operation and the inside of the pipe was allowed to fill with water through leaking joints. Down

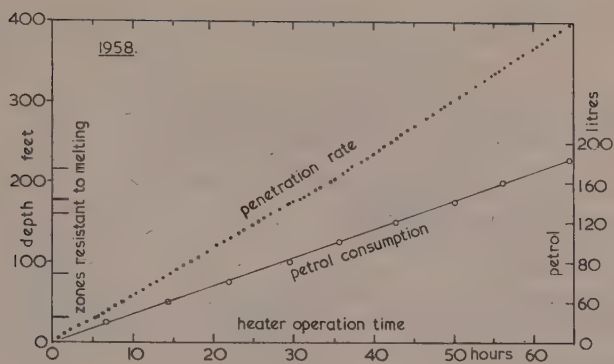


FIG. 6.

Fig. 6 — Variation of penetration and petrol consumption with heater operation time during the 1958 project.

to a depth of about 120 ft (36 m) a steady stream of air bubbles rose to the water surface whenever ice was being drilled, but below that depth very few air bubbles were seen.

The average rate of petrol consumption was about 5.0 pints per hour (2.9 litres/hour), a larger main fuel jet being used in the engine than in 1956.

4. THE 1959 DRILLING PROJECT

In the summer of 1959 two vertical unlined holes (H. 1 and H. 2) were sunk into the Austerdalsbre for the purpose of locating the bedrock and for inserting seismic explosives on the bed of the glacier. Hole H. 1 was sunk vertically at the 1958 site of the pipe mentioned above. Hole H. 2 was located at a point about 220 ft (67 m) from the steep eastern wall of the valley and on the transverse line across the glacier passing through the 1958 pipe site.

Hole H. 1 was abandoned at a nominal depth of 516 ft (157 m) on account of very slow progress in the lower 60 ft. Hole H. 2 is thought to have reached bedrock at a nominal depth of 327 ft (100 m). These depths do not include the small stretch of the supply cable due to its own weight.

The construction of the thermal drills used for boring the unlined holes is detailed in Fig. 7. The design is similar to the previous model; the same form of sheathed element is used, but the head diameter is reduced as far as possible without substantially changing the element loading. The effectively-heated length of the element sheath is about 22 inches and at a loading of 2.6 kW, the sheath dissipated about 120 watts/inch (47 watts/cm). It was realised that this design would be less efficient than the previous model, since the element is spread too close to the edge of the drill.

A union nut of stainless steel is used to close the compartment behind the element in place of the flanged coupling used previously. The arrangements for the thermostat and for the electrical connections are similar to those used in the 1958 design, but the supply cable is coupled permanently to the drill.

Two drill heads were made, one with a diameter of 3.2 inches (8.1 cm) was used down to a depth of 460 feet (140 m) in hole H. 1, and the other with a diameter of 3.38 inches (8.6 cm) was used for the remainder of hole H. 1 and for H. 2.

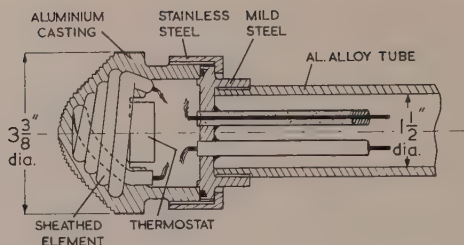


Fig. 7 — Details of the construction of the 3.38 inch (8.6 cm) diameter drill used for sinking the unlined holes H. 1 and H. 2 in 1959.

A total length of 32.5 ft (9.9 m) of aluminium-alloy tube, 1.5 inches (3.8 cm) bore, was screwed on behind the drill head. Metal discs, 3.2 inches (8.1 cm) diameter, were fitted to the tube at its mid-point and its upper end to serve as guides, and reasonably vertical holes appear to have been bored with this steering arrangement.

The records of the depth of penetration and petrol consumption for hole H. 1 are plotted against the heater operation time in Fig. 8. Throughout the operation the generator voltage was regulated at 238 volts. There is a reduction in the petrol consumption from about 2.9 to 2.5 litres per hour (5.1 to 4.4 pints/hour) due to a reduction in the size of the main jet when the depth was 393 feet (120 m). A second length of cable was connected in circuit when the drill had reached 290 feet (88 m).

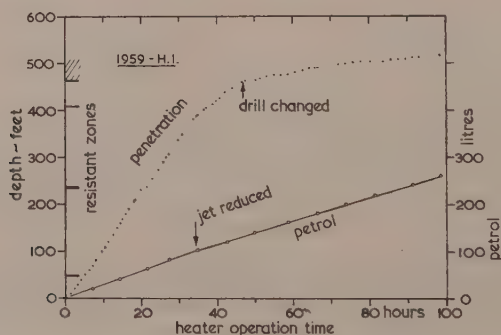


Fig. 8 — Variation of penetration and petrol consumption with heater operation time for hole H. 1 in 1959.

Apart from the few thin zones (see Fig. 8), which were obviously resistant to melting, there are several changes in the rate of penetration over long periods, which are not readily explained. About 2.6 kW were dissipated at the drill down to a depth of 290 feet (88 m), yet between 10 and 48 feet (8 and 14.5 m) depth the penetration rate is steady at about 10.9 ft/hour (3.3 m/hour), while between 70 and 230 feet (21 and 70 m) the rate is generally steady at about 12.4 ft/hour (3.8 m/hr) with the drill working at an efficiency of about 65 per cent. The efficiency remained at about this value down to a depth of about 383 feet (117 m) after the second supply cable had been connected.

The drill head was changed at a depth of 460 feet (140 m) on account of a partial short between the conductors within the compartment, which had been caused by a

small water leakage and subsequent corrosion. The resistance element remained in good condition. The leakage probably accounts for the slower rate of penetration between about 380 and 460 feet (116 and 140 m)

The very slow rate of penetration below 460 feet (140 m) is undoubtedly due to some debris retarding the melting of the ice. The boring was abandoned at a depth of about 517 feet (158 m) and when the drill was removed it was used to drill hole H. 2 without any modification.

The hole H. 2 was started for convenience 5 feet (1.5 m) below the glacier surface in the side of a sloping crevasse. The records of penetration and petrol consumption are plotted against heater operation time in Fig. 9. Apart from one thin resistant zone at 60 feet (18 m) the rate of penetration remained almost steady throughout the boring at about 9.6 ft/hour (2.9 m/hour). The generator voltage was kept close to 238 volts and the heat dissipated at the drill was about 2.6 kW throughout the boring; the drill efficiency was about 55 per cent.

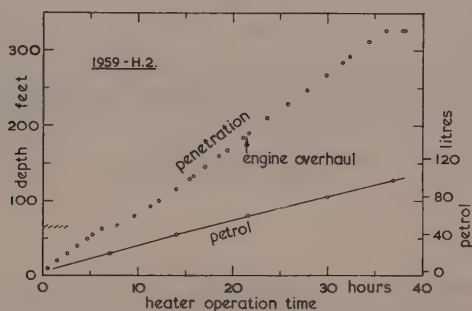


Fig. 9 — Variation of penetration and petrol consumption with heater operation time for hole H. 2 in 1959.

It was necessary to decarbonize and grind the valves of the generator engine when a depth of 160 feet had been reached. This reduced the average petrol consumption from about 2.7 to 2.5 litres/hour (4.7 to 4.3 pints/hour).

The boring came to an end in an unexpected manner. A depth of 327 feet (100 m) had been reached and the engine was stopped for refueling. Two hours later it was stopped again and refueled without any further penetration being obtained. Consequently the usual procedure of gently surging the drill up and down with the heater operating was carried out, but the drill suddenly stuck at the bottom of the hole. The heater was left operating for ten minutes without any significant penetration and surging was again possible. During this operation the alternator fuse blew and a dead short was found in the heater circuit. The hole was empty of water, and the following morning the drill was removed from the hole with some difficulty. Only the upper part of the drill head was recovered. The union nut contained a portion of the threaded part of the aluminium casting (see Fig. 7), which had melted away from the lower part.

Evidently the thermostat had not operated, but it is not clear whether this was caused by some fault within the thermostat, or to mechanical damage caused by the glacier strain, or by surging the drill. It is evident, however, that some material resistant to melting had been encountered and that the hole was sufficiently free of water for the aluminium casting to become molten as far as 3 cm above the heating element.

Identical charges of dynamite each fitted with two electric caps wired in parallel and with all connections securely soldered and taped were lowered to the bottom of

holes H. 1 and H. 2 When these came to be fired a few weeks later by Mr. D. Taylor Smith (Imperial College, London) the H. 1 charge exploded, but an open circuit was found on the leads in H. 2. This suggests exploded that the charge had become damaged, possibly for the same reason that the drill had jammed and finally fused. It therefore seems reasonable to suppose that the hole H. 2 reached bedrock.

5. FURTHER IMPROVEMENTS IN SMALL DIAMETER DRILL HEADS

The smaller efficiency of the 3.2 and 3.38 inch (8.1 and 8.6 cm) diameter drill heads is due to the comparatively large area of the leading face which is occupied by the resistance element, and by the relatively thick side walls which are necessary to form the compartment in cast aluminium. The lower efficiency of the 3.38 inch diameter drill, as compared with the 3.2 inch diameter one, arose mainly because it had thicker side walls to the compartment and conducted more heat upwards.

It would be possible to reduce the area occupied by the sheathed element by using a sheath of small diameter. If a copper casting was used in place of aluminium it would be possible to form the side walls of the compartment with a thin tube of stainless steel which could be brazed to the copper casting. Sprague and Henwood, Inc. (U.S.A.) (*) have used a copper casting with a sheathed element of the form described above, and Dr. R.L. Shreve (*) has successfully used a copper core to form another form of resistance winding; the copper was plated to prevent oxidation.

6. ACKNOWLEDGMENTS

The projects described in this paper were only made possible by the help of many Cambridge undergraduates, members of the Brathay Exploration Group, and a party from Nottingham University, who carried out back-packing and similar tasks on the glacier. Several of these people gave me special assistance in maintaining the pipe sinking operations, in particular Mr. Clive Harding, Mr. J. Tippet, Mr. R. Hull, Mr. R. White and Mr. D.W. Limbert. The help of Dr. R.L. Shreve in 1959 with his experience of similar work on the Blue Glacier was particularly valuable. To all these people I should like to express my thanks.

In addition to those mentioned in the text, I should like to record my thanks to the following firms and organisations for their kind assistance:—Norsk Braendseleolje A/S, British Insulated Callenders Cables Ltd., the Norwegian Road Authority and the Royal Norwegian Air Force.

I am particularly indebted to my expedition colleagues Mr. W.V. Lewis, Dr. J.F. Nye and Dr. J.W. Glen, for their encouragement and their scientific and technical advice. They made the task of planning and organising the recent expeditions a pleasure.

The expedition is indebted for financial help to the Royal Society, Cambridge University, the Mount Everest Foundation, Trinity College Cambridge, the Royal Geographical Society, the Tennant and Scandinavian Funds of Cambridge University and to Bristol University.

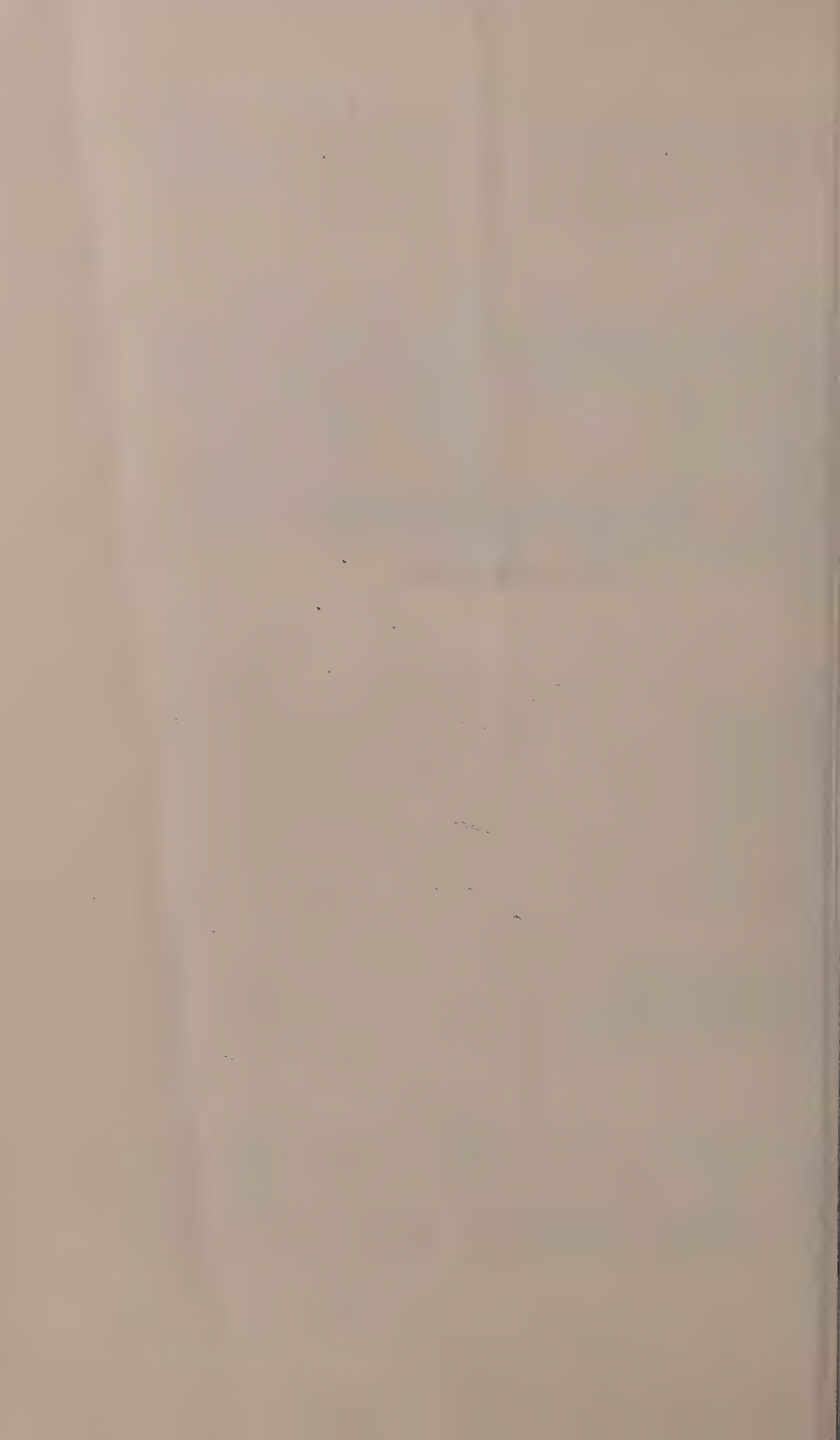
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(*) Personal communication.

MOUVEMENTS DES GLACIERS

GLACIER FLOW



STRUCTURE OF THE LOWER BLUE GLACIER, WASHINGTON ¹⁾

C.R. ALLEN, W.B. KAMB, M.F. MEIER ²⁾, and R.P. SHARP

Division of the Geological Sciences
California Institute of Technology
Pasadena, California ³⁾

SUMMARY

Structural features studied in lower Blue Glacier are the foliation pattern, an unusual longitudinal septum, ogives, crevasses, and related ice fabrics. A 300-m icefall separating the major accumulation basins from the ice tongue plays a dominant role in the formation of the principal foliated structures. Three types of ice are involved; coarse-bubbly, coarse-clear, and fine. Most and perhaps all of the fine ice represents partly recrystallized insets or infolds of firn.

An angular unconformity between stratified firn and well-foliated glacier ice is attributed to a period of high firn limits prior to 1948 and lower firn limits since.

Below the firn edge the glacier displays a foliation pattern of two sets of nested arcs, convex downglacier. These are separated by a narrow zone of strongly foliated, structurally complex ice termed the **longitudinal septum**. The foliation pattern becomes more irregular toward the terminus because of intersecting folia and discontinuities in strike.

The ogives of the Blue Glacier appear on the surface as alternating dark and white bands conformable with the arc-shaped foliation. The dark bands are underlain by well-foliated, heterogeneous material featuring unusually large percentages of fine and coarse-clear ice. The white bands are underlain by relatively massive, uniform coarse-bubbly ice.

It is inferred that the transverse foliation pattern originates in a zone of strong compressive flow immediately below the icefall. Transverse inhomogeneities created within the fall may be an important initial factor. Once formed, the foliation passively undergoes deformation within the ice tongue. The arc-shaped pattern develops within a short distance below the fall owing to differential flow. A calculation based on borehole data shows that deformation within the glacier during flow is of the correct magnitude to account for the dip of the foliation as observed at the apexes of the nested arcs, assuming that the initial attitude at the base of the icefall is essentially vertical. Complications appearing in the foliation pattern in the lower reaches of the glacier are attributed chiefly to topographic irregularities on the glacier floor, principally near the base of the icefall.

It is postulated that the longitudinal septum is formed at the base of the icefall where two ice streams, split by a large rock bastion, reunite. Differences in direction and velocity of flow at the junction produce strong compression and shear which create the intense foliation of the septum. The high content of fine ice is attributed to the inseting and infolding of firn within the icefall and in a fosse at its base.

The ogives are inferred to be primarily features formed within the icefall, but with modification in the zone of compressive flow at its base. The ogive dark bands may represent greatly compressed and partly recrystallized ice breccias which accumulated within crevasses in the icefall. There is no compelling evidence that the Blue Glacier ogives are annual features.

¹⁾ Manuscript received.

²⁾ U. S. Geological Survey, Tacoma, Washington.

³⁾ Contribution No. 964.

ICE PETROFABRIC DATA IN RELATION TO THE STRUCTURE OF BLUE GLACIER, MT. OLYMPUS WASHINGTON

W. Barclay KAMB

Division of the Geological Sciences
California Institute of Technology, Pasadena, Calif.

Ice fabrics from 14 localities on Blue Glacier show an intimate and reproducible relationship to the structure of the glacier, as outlined by the pattern of foliation in it. Only coarse bubbly ice fabrics are considered here. « Ideal » four-maximum fabrics are typical of high shear-stress situations (marginal zones), but a new type of multiple-maximum pattern, which appears to be a precursor of the 4-maximum type, has been discovered. Ogive white-band and dark-band ice, investigated at the noses of the pattern of nested arcs defined by the foliation and ogives, are well distinguished in texture and fabric. Dark band ice contains abundant fine ice inclusions and exhibits a broad maximum in c-axis orientation density, centered about the pole to the foliation plane. This pattern is unique for coarse bubbly ice from temperate glaciers, which normally shows multiple-maximum fabrics; it is similar to the fabrics of fine ice. Ogive white-band ice has complex and variable multiple-maximum fabrics indicative of a complex deformation history. Ice from the longitudinal septum of the glacier shows « ideal » 4-maximum fabrics whose orientation is not compatible with any known theory of origin of this major structural feature. With this exception the fabric data are concordant with the « structure mill » theory of foliation origin discussed separately (see paper by Allen et al.).

ZUR RHEOLOGIE VON EISSCHILDERN DER ARKTIS & ANTARKTIS

R. HAEFELI, ETH

Gletscherkommission der Schweiz. Naturforschenden Gesellschaft

ZUSAMMENFASSUNG

Unter den vereinfachenden Annahmen die der vorliegenden Theorie über den stationären Bewegungszustand von Eisschildern innerhalb des Firngebietes zu Grunde liegen, seien folgende genannt :

Allgemeine Gültigkeit des Fliessgesetzes für Eis $\dot{\epsilon} = k \cdot \tau^n$ (Glen), Gleitgeschwindigkeit auf der Sohle = 0 (Haftbedingung), horizontale Unterlage und konstante Akkumulation über die ganze Breite des Firngebietes. Nachdem die Gleichung der Firnoberfläche formuliert ist, erfolgt die Berechnung der Parameter k und n für den Fall des Inlandsis (massgebende Mittelwerte). Die Differenzen zwischen dem berechneten und dem von den Expeditions Polaires françaises (Mission P. E. Victor) gemessenen Querprofil durch das Firnggebiet des Inlandsis liegen in der Grössenordnung von höchstens 1 %. Unter den Anwendungen wird unter anderem gezeigt, dass die Höhe eines Eisschildes in Bezug auf eine Aenderung der Akkumulation sehr unempfindlich ist.

RÉSUMÉ

L'auteur expose une théorie de l'équilibre stationnaire des calottes glaciaires limitée à la zone du névé et pour l'élaboration de laquelle il admet :

- a) que la loi générale de l'écoulement de la glace (Glen) est applicable, soit $\dot{\epsilon} = k \cdot \tau^n$
- b) que la glace adhère au socle sous-glaciaire (vitesse de glissement = 0)
- c) que le taux de l'accumulation est constant et uniforme sur l'aire considérée du névé.

Après avoir établi l'équation d'équilibre de la surface, il en déduit le calcul des paramètres k et n dans le cas de l'Inlandsis du Groenland (Valeurs moyennes). Les différences entre les cotes de la surface calculées et les cotes réelles mesurées à la suite des Expéditions polaires françaises (Mission P. E. Victor) sur le profil transversal du Groenland sont d'un ordre de grandeur inférieur à 1 %.

Parmi d'autres applications de la théorie, il y a lieu de relever que l'épaisseur d'une calotte glaciaire est très peu sensible aux variations de l'accumulation.

SUMMARY

As a basis for the theory here presented of the steady-state motion of ice sheets within the firn region, the following simplifying assumptions, among others, are made : general validity of the flow law $\dot{\epsilon} = k\tau^n$ for ice (Glen), flow velocity at the bed vanishes (no slip on the bed), horizontal bed, and constant accumulation over the entire firn region. Formulation of the equation for the firn surface is followed by calculation of the parameters k and n for the case of the Greenland ice cap (adequate average values). The calculated surface profile differs from the profile measured by the Expeditions Polaires Françaises (mission of P. E. Victor) across the firn region of the Greenland ice cap by at most 1 %. Among the applications of the theory, it is shown that the height of an ice sheet is very insensitive to changes in accumulation.

1. PROBLEMSTELLUNG

Der nachstehende Versuch, den Gleichgewichtszustand und insbesondere die Gleichung der Firnoberfläche eines stationären Eisschildes näherungsweise zu formulieren, wurde vor allem durch folgende Arbeiten angeregt : Die experimentellen und theoretischen Studien englischer und schweizerischer Physiker und Glaziologen [1-5], die die Ergebnisse der französischen Groenlandexpeditionen von P. E. Victor [6] u[7].

Als Hauptproblem wird zunächst der ebene, stationäre Bewegungszustand

und die Gestaltung der Oberflächenform im Firngebiet eines streifenförmigen Eisschildes bei einer konstanten mittleren Akkumulation behandelt. Die gesuchte Funktion lautet :

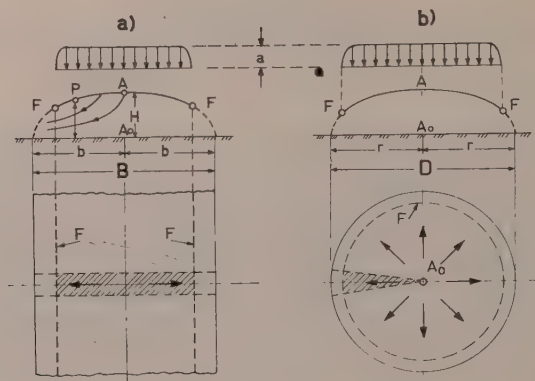


Fig. 1 — a) Streifenförmiger Eisschild b) Spärischer Eisschild a = Akkumulation pro Jahr.

(1) $y = \varnothing (x, H, a, k, n)$ worin nach Fig. 1 bedeuten :

y = Eismächtigkeit eines im Abstand x von Eismitte A gelegenen Punktes F der Firnoberfläche.

H = grösste Eismächtigkeit des Eisschildes in A .

a = Jahresakkumulation (Dicke der pro Jahr gebildeten Eisschicht für $\rho = 917 \text{ kg/m}^3$).

k, n = Eisparameter (vgl. Gl. 2)

Nachdem auf Grund einer Bilanzbetrachtung (Kontinuitätsgleichung) und des Fließgesetzes von Glen [1] die allgemeine Gleichung (1) der Firnoberfläche des streifenförmigen Eisschildes aufgestellt ist, gilt es, die theoretische Oberfläche $A-C$ gemäss Fig. 1a mit dem von den Expeditions Polaires Françaises vermessenen Oberflächenprofil zu vergleichen. Ausgehend vom Kulminationspunkt A und einem weiteren bergseits der Firnlinie gelegenen Punkt C , dessen Koordinaten c und h bekannt sind, werden die Konstanten des Eises (k und n) bei bekannter Akkumulation a so gewählt, dass eine möglichst vollkommene Übereinstimmung zwischen dem gemessenen und berechneten Profil der Firnoberfläche erzielt wird. (vgl. Fig. 4).

Anschliessend wird in analoger Weise die Gleichung für die Oberfläche des sphärischen Eisschildes (Fig. 1b) formuliert, deren Differentialgleichung sich von derjenigen des streifenförmigen Eisschildes nur durch einen Formfaktor unterscheidet. Ferner werden gewisse allgemeine Zusammenhänge zwischen der Akkumulation a und der grössten Eismächtigkeit H formuliert und an Hand von Beispielen erläutert.

2. EXPERIMENTELLE UND THEORETISCHE GRUNDLAGEN

Als wichtigste experimentelle Grundlage, die sich bei zahlreichen neueren Untersuchungen bewahrt hat [23] und für das polykristalline Gletschereis als gute Näherungslösung betrachtet werden darf, benützen wir das an polykristallinem Eis Fig. 1 (selon U. Monterin, Boll. Comitato Glaciologico Italiano, 1932).

$$(2) \quad \frac{dv_x}{dz} = \varepsilon = k \tau^n \quad (\text{vgl. Fig. 2})$$

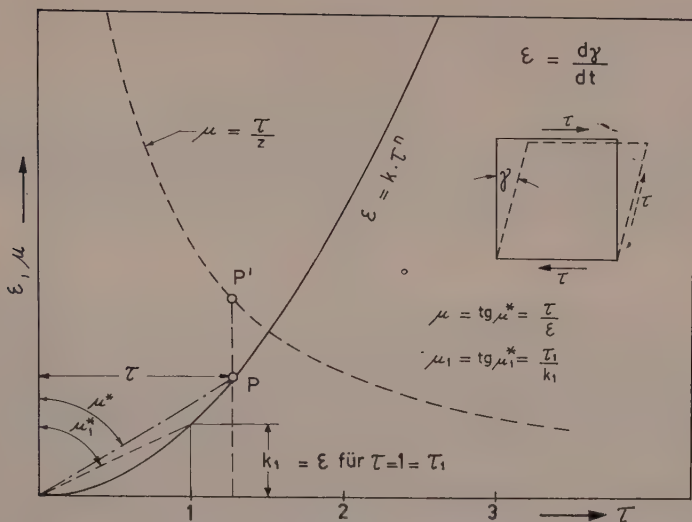


Fig. 2 — Fliesskurve von Eis; μ = scheinbare Zähigkeit.

worin der Parameter k hauptsächlich von der Temperatur des Eises [8] und der Exponent n u.a. von den kristallographischen Verhältnissen sowie der Verformungsvorgeschichte der betreffenden Eisart abhängig ist [5]. Um für den Temperaturparameter eine leicht vorstellbare Dimension zu erhalten, schreiben wir Gl. 2 in folgender Form :

$$(3) \quad \varepsilon = k_1 \left(\frac{\tau}{\tau_1} \right)^n \quad \text{worin } \tau_1 = \text{Einheit der Spannung}$$

Für $\tau = \tau_1$ wird $\varepsilon = k_1$, wonach der Parameter k_1 definiert werden kann als diejenige Verformungsgeschwindigkeit (Winkelgeschwindigkeit in sec^{-1}), die sich unter der Einheit τ_1 der Schubspannung einstellt. Je tiefer die Eistemperatur, umso kleiner ist k bzw. k_1 . Die Beziehungen zwischen den Parametern k und k_1 , die sich nur in ihrer Dimension unterscheiden, lauten :

$$k = \frac{k_1}{\tau_1^n}; \quad k_1 = k \cdot \tau_1^n$$

Um den Charakter der Fliesskurve des Eises als Ausdruck einer unvollkommenen zähen Flüssigkeit zu betonen und den Vergleich mit einer Newton'schen zähen Flüssigkeit zu erleichtern, ist es zweckmässig, die «scheinbare Zähigkeit» μ einzuführen, die wir auf Grund von Gl. 2-4 und Fig. 2 wie folgt definieren :

$$(5) \quad \mu = \text{tg } \mu^* = \frac{\tau}{\varepsilon} = \frac{\tau^{(1-n)}}{k} = \frac{\tau^{(1-n)} \cdot \tau_1^n}{k_1}$$

$$\text{Für } \tau = \tau_1 : \mu_1 = \text{tg } \mu_1^* = \frac{\tau_1}{k_1}$$

$$\text{Für: } \tau = \tau_2 : \mu = \operatorname{tg} \mu_2^* = \frac{\tau_2^{(1-n)}}{k_1} \cdot \tau_1^n$$

$$(5a) \quad \frac{\mu_2}{\mu_1} = \left(\frac{\tau_2}{\tau_1} \right)^{(1-n)}$$

Für den Spezialfall $n = 1$ wird die scheinbare Zähigkeit μ , die als Tangens des Winkels μ^* definiert werden kann, welchen der Leitstrahl r der Fließkurve mit der Lotrechten einschliesst, identisch mit der Newton'schen Zähigkeit η , d.h.

$$\mu_{(n=1)} = \frac{\tau_1}{k_1} = \eta$$

Da die Temperatur des Eisschildes von Punkt zu Punkt ändert, so müsste streng genommen mit einem ortsveränderlichen k_1 -Wert gerechnet werden [3,8]. Der Umstand, dass sowohl die Eistemperatur wie auch die Scherspannung längs einer Vertikalen mit der Tiefe in der Regel zunehmen und an der Gletschersohle ihr Maximum erreichen, bewirkt eine starke Krümmung des in Fig. 3 dargestellten Profils der horizontalen Geschwindigkeiten in seinem unteren Teil, auf die schon Nye aufmerksam gemacht hat [9]. Wird einfachheitshalber mit einem konstanten mittleren k -Wert gerechnet, so kann dieser Fehler, wie später noch gezeigt wird, durch einen relativ hohen n -Wert kompensiert werden. (Das in Fig. 3 dargestellte Geschwindigkeitsprofil entspricht z.B. dem Wert $n = 4$).

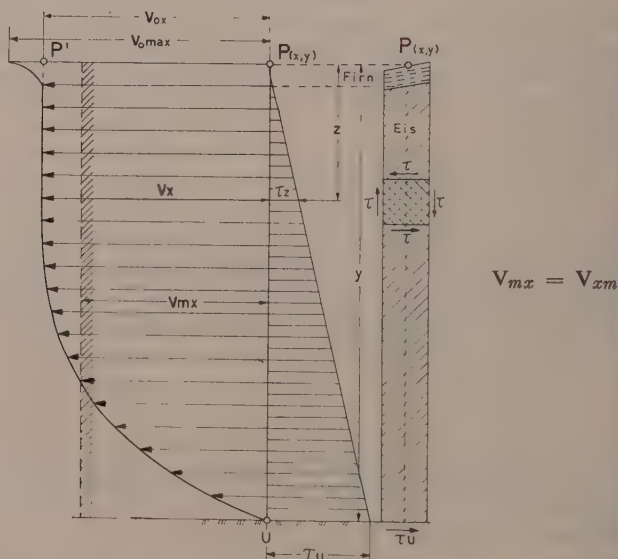


Fig. 3 — Geschwindigkeitsprofil (horizontale Komponenten).

Man beachte ferner, dass das Geschwindigkeitsprofil infolge der wesentlich geringeren Zähigkeit der oberflächlichen Firnschichten theoretisch eine kleine Anomalie in Form einer lokalen Spitze aufweisen dürfte, wie sie auch in der Schneedecke sowie in alpinen Firngebietern häufig beobachtet [10] und zurzeit von der Internationalen glaziologischen Grönlandexpedition auf dem Inlandeis näher untersucht wird [11].

Dank der im Vergleich zur totalen Eismächtigkeit relativ geringen Dicke der Firnschicht wird dieser evt. Anomalie in den Bilanzbetrachtungen nur insofern Rechnung getragen, als man bei der Inhaltsberechnung des Geschwindigkeitsprofils nicht von der reellen Oberflächengeschwindigkeit $v_{x \max}$ ausgeht, sondern von der theoretischen Geschwindigkeit v_0 , die sich normalerweise einstellen würde, wenn die Firndecke durch Eis ersetzt würde. Der in Fig. 3 angegebene Wert v_{xm} bedeutet dann die mittlere horizontale Geschwindigkeitskomponente des durch Pfeile markierten lotrechten Geschwindigkeitsprofils $P-P'-U$ gemäss der Gleichung :

$$(6) \quad v_{xm} = \int_0^y \frac{v_x \cdot dz}{y}$$

3. BERECHNUNGSANNAHMEN

Der nachstehend entwickelten Theorie liegen folgende Annahmen zu Grunde :

1) Das Fliessgesetz von Glen (Gl. 2) gelte unabhängig von der Grösse des hydrostatischen Ueberlagerungsdruckes, was von G.P. Rigsby [12] bis zu Drücken von 350 kg/cm² experimentell nachgewiesen wurde.

2) Es handle sich um einen laminaren, stationären Strömungsvorgang (bruchlose Verformung), der sich unter dem Einfluss der Schwerkraft und der inneren Reibung so langsam vollzieht, dass die Beschleunigungskräfte vernachlässigbar sind (schleichende Bewegung).

3) Innerhalb des betrachteten Firngebietes $C-A-C$, für welches die Gleichung der Oberfläche zu berechnen ist, wird die mittlere Akkumulation a als konstant vorausgesetzt.

4) Innerhalb des ganzen Firngebietes $F-A_0-F$ sei die Haftbedingung erfüllt, d.h. das Eis gleite nicht auf seiner Unterlage ($v_u = 0$). Dabei ist es in diesem Zusammenhang belanglos, ob diese Haftung durch das Festfrieren des Eises an der Gletschersohle (Permafrost), durch eine grosse Rauigkeit der Kontaktfläche, relativ kleine Schubspannungen oder durch das Zusammenwirken von 2 oder 3 dieser Einflüsse bedingt ist. Auf Grund der Untersuchungen von Robin [3] und Holtz

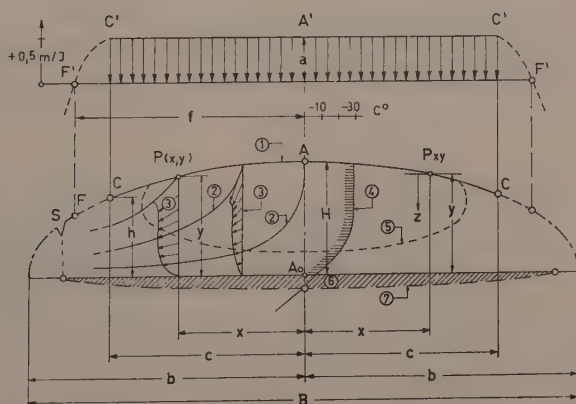


Fig. 4 — a = Akkumulation in m Eis/Jahr — c = Halbe Breite des Firngebietes mit mittlerer Akkumulation a — f = Halbe Breite des Firngebietes.
 F = Firnlinie — b = Halbachse der elliptischen Kurve (1) — S = Spaltenzone.
 (1) = Firnoberfläche (2) = Laufkurven (3) = Geschwindigkeitsprofil (4) = Eistemperatur in Eismitte (5) = Isothermenverlauf (6) = Permafros (7) = Druckschmelzpunkt.

7) Sämtliche Formulierungen beziehen sich nur auf die horizontalen Komponenten der Geschwindigkeiten, die für die Bilanzbetrachtung massgebend sind. Ihre Gültigkeit beschränkt sich auf das Firngebiet.

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$$(9) \text{ Nach Gl. 2-3 : } \frac{dv_x}{dz} \cong k \cdot \tau^n = k_1 \cdot \left(\frac{\tau}{\tau_1} \right)^n$$

$$(10) \text{ Mittlere Profilgeschwindigkeit : } v_{xm} = \frac{a \cdot x}{y} \text{ (Bilanzgleichung)}$$

Gl. 10 ist die dem stationären Zustand des Eisschildes entsprechende Kontinuitätsbedingung.

4.2. Geschwindigkeitsverteilung längs einer Lotrechten (Vertikales Geschwindigkeitsprofil)

Aus den Gleichungen 8 und 9 folgt (vgl. Fig. 5) :

$$\frac{dv_x}{dz} = k \cdot [\varrho \cdot \operatorname{tg} \alpha \cdot z]^n$$

$$v_x = k \cdot \varrho^n \cdot \operatorname{tg} \alpha^n \cdot \int_0^y z^n \cdot dz$$

$$(11) \quad v_x = \frac{k \cdot \varrho^n \cdot \operatorname{tg} \alpha^n}{n+1} [y^{n+1} - z^{n+1}] = C_1 [y^{n+1} - z^{n+1}]$$

$$(12) \quad v_0 = \frac{k \cdot \varrho^n \cdot \operatorname{tg} \alpha^n}{n+1} \cdot y^{n+1} = C_1 \cdot y^{n+1} ; C_1 = \frac{k \cdot \varrho^n \cdot \operatorname{tg} \alpha^n}{n+1}$$

Die mittlere Profilgeschwindigkeit v_{xm} berechnet sich somit zu :

$$v_{xm} = \int_0^y \frac{v_x \cdot dz}{y} = \frac{C_1}{y} \cdot \int_0^y (y^{n+1} - z^{n+1}) \cdot dz$$

$$= C_1 \cdot \frac{n+1}{n+2} \cdot y^{n+1}$$

$$(13) \quad v_{xm} = C_2 \cdot y^{n+1} ; C_2 = C_1 \cdot \frac{n+1}{n+2} = \frac{k \cdot \varrho^n \cdot \operatorname{tg} \alpha^n}{n+2}$$

$$(13a) \quad \frac{v_0}{v_{xm}} = \frac{n+2}{n+1}$$

4.3. Differentialgleichung der Firnoberfläche

Aus den Gleichungen 7, 10 und 13 ergibt sich folgende Differentialgleichung für die Kurve der Firnoberfläche :

$$\begin{aligned} v_{xm} = \frac{a \cdot x}{y} &= - \frac{k \cdot \varrho^n}{n+2} \cdot \left(\frac{-dy}{dx} \right)^n \cdot y^{n+1} \\ &= - C_3 \cdot \left(\frac{-dy}{dx} \right)^n \cdot y^{n+1} \end{aligned}$$

$$(14) \quad \frac{dy}{dx} = - \sqrt[n]{\frac{a}{C_3}} \cdot \frac{x^{1/n}}{y^{(n+2)/n}} = \operatorname{tg} \alpha ; C_3 = \frac{k \cdot \varrho^n}{n+2}$$

4.4. Gleichung der Kurve A-C der Firnoberfläche

Aus Gl. 14 folgt :

$$y^{(n+2)/n} \cdot dy = - \sqrt[n]{\frac{a}{C_3}} \cdot x^{1/n} \cdot dx$$

Die Integration dieser Differentialgleichung ergibt folgenden Ausdruck für die Eismächtigkeit y in Funktion des Abstandes x von Eismitte :

$$\begin{aligned} \frac{y^{2(n+1)/n}}{2(n+1)/n} &= - \sqrt[n]{\frac{a}{C_3}} \cdot \frac{x^{n+1/n}}{n+1/n} + C_4 \\ \text{für } x = 0; y = H : C_4 &= \frac{H^{2(n+1)/n}}{2(n+1)/n} \\ y &= \left[H^{2(n+1)/n} - 2 \sqrt[n]{\frac{a}{C_3}} \cdot x^{n+1/n} \right]^{n/2(n+1)} \\ (15) \quad y &= [H^{2(n+1)/n} - C_5 \cdot x^{n+1/n}]^{n/2(n+1)}; C_5 = \frac{2}{\varrho} \sqrt[n]{\frac{a(n+2)}{k}} \end{aligned}$$

Wie Gl. 15 zeigt, handelt es sich hier um eine geschlossene Kurve von elliptischem Charakter, mit der kleinen Halbachse $= H$, der grossen Halbachse b , die sich für $y = 0$ (15) wie folgt berechnet :

$$(16) \quad b = \frac{H^2}{\sqrt[n+1]{\left(\frac{2}{\varrho}\right)^n \cdot \frac{a(n+2)}{k}}} \quad (\text{vgl. Fig. 4 und 5})$$

Man beachte dabei, dass diese Halbachse b wie auch das Kurvenstück C-F-T nur eine mathematische, aber keine reelle Bedeutung haben. Deshalb braucht b nicht mit der halben Breite des Eisschildes übereinzustimmen. Ferner ist zu beachten, dass die Kurve für y durch eine horizontale Tangente in A (Kulmination) und eine vertikale Tangente in T gekennzeichnet ist. Die erstgenannte ist reell, da die Firnoberfläche der Kulmination A stetig verläuft.

4.5. Bestimmung der Parameter n und k

Sind bei der Untersuchung eines bestimmten Eisschildes ausser dem Kulminationspunkt A zwei weitere Punkte der Firnoberfläche durch ihre Koordinaten (x, y) gegeben, so können auf Grund von Gl. 15 die für den betreffenden Fall massgebenden Mittelwerte von k und n berechnet werden.

Lässt man zunächst n offen und berechnet k in Funktion von n , indem man ausser A nur den Randpunkt C mit den Koordinaten c und h (vgl. Fig. 4, 5 und 6) als bekannt voraussetzt, so ergibt sich aus Gl. 15 folgender Wert für k :

$$(17) \quad k = \frac{2^n(n+2) \cdot a \cdot c^{n+1}}{\varrho^n [H^{2(n+1)/n} - h^{2(n+1)/n}]^n}; k_1 = k \cdot \tau_1^n(4)$$

Man kann nun n solange variieren, bis die gemessene Firnoberfläche des Eisschildes (Fall a) mit der nach Gl. 15 berechneten möglichst gut übereinstimmt.

Es besteht andererseits die Möglichkeit, den Parameter k zu eliminieren, indem man dessen Wert (Gl. 17) in Gl. 15 und 16 einführt. Dabei ergeben sich folgende Ausdrücke :

$$(18) \qquad \frac{a}{k} = Q^n \cdot \frac{[H^{2(n+1)/n} - h^{2(n+1)/n}]n}{2^n(n+2) \cdot C^{n+1}}$$

$$(19) \qquad y = \frac{\left\{ 1 - \left[1 - \left(\frac{h}{H} \right)^{2(n+1)/n} \right] \cdot \left(\frac{x}{c} \right)^{n+1/n} \right\}^{n/2(n+1)} \cdot H}{\left\{ 1 - N \left(\frac{x}{c} \right)^{n+1/n} \right\}^{n/2(n+1)} \cdot H ; N = 1 - \left(\frac{h}{H} \right)^{2(n+1)/n}}$$

$$(20) \qquad \text{für } y = 0 : x = b = \frac{c}{N^{n/n+1}} \qquad b = \text{grosse Halbachse}$$

Man beachte, dass in diesem Falle y (19) und b (20) bei gegebenen Koordinaten der Punkte A und C allein vom Parameter n abhängen.

5. ANWENDUNG AUF DAS GRÖNLÄNDISCHE INLANDSIS

Das von den Expeditions Polaires Françaises aufgenommene Querprofil [6,24], das 1959 von der Internat. glaziologischen Grönlandexpedition erneut vermessen wurde ist in Fig. 6 dargestellt. Gemäss Kap. IV/5 gehen wir beim Vergleich zwischen der gemessenen und berechneten Firnoberfläche wie folgt vor :

Ausser dem Kulminationspunkt A mit $H = 3160$ m wählen wir im gemessenen Querprofil (Fig. 6) den auf 2000 m oberhalb der Firnlinie liegenden Endpunkt C unserer Funktion y . Die Entfernung dieses Punktes von A beträgt $c = 385$ km.

Als mittlere Akkumulation zwischen den Punkten A und C wurde nach Diamond [14] (vgl. Fig. 6, Kurve 11) 45 cm Wasser bzw. 50 cm Eis ($a = 0,50$ m pro Jahr) geschätzt, wobei die Akkumulation im westlichen Teil grösser ist als im östlichen. Sorge hat in Station Eismitte auf Grund der Firnschichten eine mittlere Akkumulation von 31,4 cm (H₂O) festgestellt [15,16].

Für Eis verschiedenen für n angenommenen Werte ($n = 1, 2, 3$ und 4) und $a = 0,50$ m Eis/Jahr berechnen sich auf Grund der Gleichungen 5, 17, 19 und 20 folgende Beiwerte :

TABELLE 1
Beiwerte für verschiedene n -Werte (k_1, μ_1, μ_2, N, b)

n Einheiten \rightarrow	k_1 Gl. 17 sec ⁻¹	$\tau_1 = 1 \text{ kg/m}^2$ μ_1 Gl. 5e Poise	$\tau_2 = 0,5 \text{ kg/cm}^2$ μ_2 Gl. 5a Poise	N Gl. 19	b Gl. 20 km
$n = 1$	$0.184 \cdot 10^{-12}$	$5,4 \cdot 10^{14}$	$5,4 \cdot 10^{14}$	0,8394	420,3
2	$0.314 \cdot 10^{-16}$	$3,2 \cdot 10^{18}$	$6,4 \cdot 10^{14}$	0,7458	468,2
3	$0.505 \cdot 10^{-20}$	$2,0 \cdot 10^{22}$	$8,0 \cdot 10^{14}$	0,7046	500,6
4	$0.84 \cdot 10^{-24}$	$1,2 \cdot 10^{20}$	$9,5 \cdot 10^{14}$	0,6812	523,5

Fig. 6

- ① Verlauf der Firnoberfläche für $n = 1$.
- ② Verlauf der Firnoberfläche für $n = 2$.
- ③ Gemessene Firnoberfläche (EPF).
- ④ Verlauf der Firnoberfläche für $n = 4$.
- ⑤ Mittlere Profilgeschwindigkeiten (Horizontalkomponenten m/Jahr).
- ⑥ Neigung der Firnoberfläche in % ($n = 4$).
- ⑦ Scherspannung τ_n an der Basis (kg/cm^2) für $n = 4$.
- ⑧ Vertikales Geschwindigkeitsprofil (Horizontalkomp. m/Jahr, $n = 4$).
- ⑨ Verteilung der Scherspannung im Vertikalschnitt (kg/cm^2) $n = 4$.
- ⑩ Konstante Akkumulation (A-C) $a = 0,5$ m/Jahr. (Eis)
- ⑪ Verteilung der Akkumulation nach Diamond in m $\text{H}_2\text{O/Jahr}$.
- ⑫ Theoretische Basisebene.
- ⑬ Seismisch aufgenommene Basis nach Holtzscherer (EPF, Mission P.E. Victor).
- ⑭ Gravimetrisch aufgenommene Basisfläche nach Holtzscherer (EPF), [6].

TABELLE 2

Vergleich der berechneten und gemessenen Ordinaten y in m
($y = \text{Höhe in m.ü.M.}$, vgl. Fig. 6).

Kurve no. →	Berechnete Werte y			Gemessen ¹⁾ (ca.) (5) m.ü.M.	Differenz (4-5)		
	(1)	(2)	(4)		Δy_{4-5}	$\frac{\Delta y_{4-5}}{y_5}$	
	$n \rightarrow$	$n = 1$	$n = 2$				$n = 4$
Pkt. Abszisse $A \quad 0$	3160	3160	$y $ 3160	y^{I} 3160	m —	% —	
P_1	50	3150	3125	3100	3110	—10	—0,3
	77	3119	3093	3043	3055	—12	—0,4
P_2	100	3115	3050	3000	3020	—20	—0,7
	150	3055	2955	2885	2910	—25	—0,9
	154	3054	2946	2865	2895	—30	—1,0
P_3	200	2955	2835	2752	2760	— 8	—0,3
	231	2888	2728	2644	2660	—16	—0,6
P_4	250	2830	2680	2572	2580	— 8	—0,3
	300	2645	2490	2390	2410	—20	—0,8
	308	2606	2450	2362	2360	+ 2	+0,1
	350	2350	2240	2182	2170	+12	+0,6
	385	2000	2000	2000	2000	—	—
1) Aus dem Plan der EPF entnommen [6]				Mittel	—12,3	—0,5	

Während sich die scheinbare Zähigkeit μ_1 , auf die Einheit der Schubspannung (1 kg/m^2), d.h. auf eine sehr kleine Spannung bezieht, bedeutet μ_2 die scheinbare Zähigkeit für $\tau = 5000 \text{ kg/m}^2 = 0,5 \text{ kg/cm}^2$, was ungefähr einer mittleren Scherspannung des Eisschildes entspricht (vgl. Fig. 6). Für $n = 1 - 4$ resultieren dabei scheinbare Zähigkeiten μ_2 , deren Grössenordnung den früher ermittelten, mittleren Zähigkeiten μ_m des grönländischen Eisschildes wie auch der Eiskalotte auf Jungfraujoch entsprechen (vgl. [7], Tabelle 2, p. 629).

In Tabelle 2 werden die für verschiedene n -Werte auf Grund von Gl. 19 und der in Tabelle 1 enthaltenen Hilfswerte N berechneten Ordinaten y zusammengestellt und mit den gemessenen Oberflächenkoten verglichen (y_5).

Profil des Inlandeises (Exp. Polaires Françaises).

Aus dem in Fig. 6 und Tabelle 2 dargestellten Vergleich zwischen der für $n = 4$ berechneten Oberflächenkurve (4) einerseits und der von den EPF gemessenen Firnoberfläche (3) andererseits geht hervor, dass die Differenz zwischen den beiden Kurven weniger als 1% der Eismächtigkeit beträgt. Durchschnittlich liegen die analytisch ermittelten Punkte nur ca. 0,5% von y tiefer als die gemessenen, während die für $n = 2$ berechnete Kurve wesentlich höher liegt. Die beste Uebereinstimmung zwischen Messung und Rechnung würde man für einen n -Wert zwischen 3 und 4 — jedoch näher bei 4 — erhalten.

Man beachte, dass die für $n = 4$ berechnete und in Fig. 6 dargestellte Oberflächenkurve (4) auch noch unterhalb des Punktes C, wo ihr streng genommen keine reelle Bedeutung mehr zukommt, relativ gut übereinstimmt. Ferner ist der sehr grosse Unterschied zwischen der Lösung für eine Newton'sche zähe Flüssigkeit (Kurve 1 für $n = 1$) und der gemessenen (3) bzw. für $n = 4$ berechneten Oberfläche (4) bemerkenswert.

Für die Oberflächenneigungen (Fig. 6, Kurve 6), die Schubspannungen τ_u (7) an der Sohle sowie die mittleren Profilgeschwindigkeiten v_{xm} (5) ergeben sich folgende Werte :

TABELLE 3
Zusammenstellung von $tg \alpha$, τ_u , v_{xm} , α , v_{0x} (für $n = 4$)

Punkt (Fig. 6)	x km	y m	$tg \alpha$ %	τ_u kg/cm ²	v_{xm} m/Jahr	v_{0x} m/J
A	0	3160	0	0	0	0
P ₁	77	3043	0,190	0,53	12,6	15,2
P ₂	154	2865	0,246	0,65	27	32
P ₃	231	2644	0,310	0,75	44	53
P ₄	308	2362	0,400	0,86	65	78
C	385	2000	0,540	0,98	96	115

6. GRUNDGLEICHUNGEN FÜR DEN SPHÄRISCHEN EISSCHILD

Die Ableitung der Oberflächengleichung für den sphärischen Eisschild erfolgt analog wie beim streifenförmigen, wobei als einziger Unterschied die Kontinuitäts- oder Bilanzgleichung zu nennen ist, die hier wie folgt lautet ($x = r = \text{Radiusvektor}$) :

$$a \cdot x^2 \cdot \pi = 2 \cdot x \cdot \pi \cdot y \cdot v_{xm}$$

$$(10a) \quad v_{xm} = \frac{a \cdot x}{2y} \quad (\text{statt } \frac{ax}{y} \text{ nach Gl. 10})$$

Die entsprechenden Grundgleichungen lauten :

$$(15a) \quad y = \left[H^{2(n+1)/n} \cdot \frac{2}{\varrho} \cdot \sqrt{\frac{a(n+2)}{2k}} \cdot x^{(n+1)/n} \right]^{n/2(n+1)}; C_5^* = \frac{2}{\varrho} \cdot \sqrt[n]{\frac{a(n+2)}{2k}}$$

$$(16a) \quad r = \frac{H^2}{\sqrt[n+1]{\left[\frac{2}{\varrho} \right]^n \cdot \frac{a(n+2)}{2k}}} = \text{grosse Halbaxe}$$

$$(17a) \quad k = \frac{2^{n-1}(n+2) \cdot a \cdot c^{n+1}}{\varrho^n [H^{2(n+1)/n} - h^{2(n+1)/n}]^n}; k_1 = k \cdot \tau_1^n$$

$$(19a) \quad y = \left\{ 1 - N \left(\frac{x}{c} \right)^{(n+1)/n} \right\}^{n/2(n+1)} \cdot H$$

$$(20a) \quad r = \frac{c}{N^{n/(n+1)}}; N = 1 - \left(\frac{h}{H} \right)^{2(n+1)/n}$$

7. ANWENDUNGEN

Aus den verschiedenen Anwendungsgebieten werden nachstehend folgende zwei Beispiele herausgegriffen :

7.1. Berechnung der mittleren Akkumulation eines im stationären Zustand befindlichen Eisschildes.

Nach Gl. 17 bzw. 17a kann a berechnet werden, wenn wir den ungefähren Verlauf der Firnoberfläche kennen (H, h & c) und ausserdem die Werte k und n . Für den sphärischen Eisschild erhält man z.B. aus Gl. 17a :

$$(21) \quad a = \frac{\varrho^n [H^{2(n+1)/n} - h^{2(n+1)/n}]}{2^{(n-1)}(n+2) \cdot C^{n+1}} \cdot k; k = \frac{k_1}{\tau_1^n}$$

Als Zahlenbeispiel wählen wir einen Eisschild von ähnlichen Dimensionen wie diejenigen der Antarktis und setzen dabei für k und n die bei der Analyse des grönländischen Inlandeises (Profil EPF) erhaltenen Wertepaare ein (z.B. $n = 3$). Liegt die Firnlinie nur wenig über dem Meeresspiegel, so kann h gegenüber H vernachlässigt werden, wonach sich Gl. 21 wie folgt vereinfacht :

$$(22) \quad a \sim \frac{\varrho^n \cdot H^{2(n+1)}}{2^{(n-1)}(n+2) \cdot c^{n+1}} \cdot k$$

Setzt man für $\varrho = 900 \text{ kg/m}^3$, $H = 4500 \text{ m}$, $c = 2000 \text{ km}$, $\tau_1 = 1 \text{ kg/m}^2$, $n = 3$ und $k_1 = 0,505 \cdot 10^{-20} \cdot \text{sec}^{-1}$ (Tab. 1), so folgt :

$$\begin{aligned} a &= \frac{\varrho^3 \cdot H^8}{20 \cdot c^4} \cdot k = \frac{0.9^3 \cdot 10^9 \cdot 4.5^8 \cdot 10^{24}}{20 \cdot 2^4 \cdot 10^{24}} \cdot k \\ &= 0.38 \cdot 10^{12} \cdot 0.505 \cdot 10^{-20} = 1.92 \cdot 10^{-9} \text{ m/sec} \\ &= 6 \text{ cm Eis/Jahr} = 5,4 \text{ cm H}_2\text{O/Jahr} \end{aligned}$$

Die Frage, ob auf diesem indirekten Wege ein Beitrag zur Lösung des Akkumulationsproblems der Antarktis geleistet werden kann, bleibt offen. Es ist klar, dass eine Uebertragung der in der Arktis und insbesondere im grönländischen Inlandeis gewonnenen Erfahrungen auf die Antarktis nur mit der grössten Vorsicht und unter Kenntnis der andersgearteten Verhältnisse erfolgen darf. Im engen Rahmen dieser Studie kann es sich nur darum handeln, auf gewisse allgemeine Zusammenhänge zwischen der Bilanzbetrachtung einerseits und den rheologischen Gesetzmässigkeiten andererseits hinzuweisen [17-20].

7.2. Verhältnis zwischen der Akkumulation a und der grössten Eismächtigkeit im stationären Zustand eines Eisschildes

Bei gleichen Werten n und k folgt aus Gl. 22 : (für $h \sim 0$) :

$$\frac{a_1}{a_2} \cong \left(\frac{H_1}{H_2} \right)^{2(n+1)} \quad \& \quad \text{für } n = 4 : \frac{a_1}{a_2} \cong \left(\frac{H_1}{H_2} \right)^{10}$$

$$(23) \quad n = 4 : \frac{H_2}{H_1} \cong \sqrt[10]{\frac{a_2}{a_1}}; n = 3 : \frac{H_2}{H_1} \cong \sqrt[8]{\frac{a_2}{a_1}}$$

Daraus ergibt sich, dass die Eismächtigkeit H sehr unempfindlich ist in Bezug auf eine Aenderung der Akkumulation, und dies umso mehr, je höher der n -Wert liegt. Wäre z.B. $a = 2a_1$, d.h. würde sich die Akkumulation auf den doppelten Betrag erhöhen, ohne dass k ändert, so würde theoretisch eine Zunahme der grössten Eismächtigkeit bei $n = 4$ um nur 7,8% und für $n = 3$ um ca. 9% gemügen, damit wieder ein stationärer Zustand eintreten könnte. Für $n = 1$, d.h. für eine Newton'sche zähe Flüssigkeit, würde der entsprechende Prozentsatz dagegen 19% betragen, d.h. die Eismächtigkeit würde dann auf eine Aenderung von a etwas empfindlicher reagieren. Frägt man andererseits nach dem prozentualen Höhenzuwachs, welche ein Eisschild von den ungefähren Dimensionen der Antarktis erfahren würde, wenn seine Akkumulation auf den 10-fachen Betrag ansteigen sollte (z.B. von 10 cm auf 100 cm Eisdiche pro Jahr), so ergibt sich für $n = 4$ und den entsprechenden k -Wert nur ein Betrag von 26% bzw. von ca. 33% für $n = 3$. *Es ist deshuld nicht sehr wahrscheinlich dass das Innere der Antarktis a des Inlandsis während der Eiszeit wesentlich anders ausgesehen hat als heute.*

8. SCHLUSSBEMERKUNGEN

Die dargelegte Theorie über das Grossrelief und die Geschwindigkeitsverhältnisse im Firngebiet eines Eisschildes wurde vor allem im Hinblick auf die internationale glaziologische Grönlandexpedition als Arbeitshypothese für den rheologischen Teil des Expeditionsprogramms entwickelt [21]. Sie diente z.T. auch dazu, um die für die geodätischen Arbeiten erforderlichen Messgenauigkeiten zu ermitteln. Eine abschliessende Beurteilung der Brauchbarkeit dieser theoretischen Ansätze dürfte erst auf Grund der ausgewerteten Resultate (*) der zur Zeit noch im Gange befindlichen Expedition und den Ergebnissen weiterer Expeditionen der Arktis und Antarktis möglich sein.

Die vorliegende Studie wurde ungefähr gleichzeitig mit einer sehr interessanten Untersuchung von J.F. Nye ausgearbeitet, die ein ähnliches Thema behandelt [9] [23]. Trotzdem die beiden unabhängig von einander entstandenen Arbeiten von ganz verschiedenen Voraussetzungen ausgehen, gelangen sie hinsichtlich der auffallenden Unempfindlichkeit der Eismächtigkeit in Bezug auf die Akkumulation zu ähnlichen Ergebnissen.

Zürich, den 23. April 1960.

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MEASUREMENT OF THE STRAIN OF A GLACIER SNOUT

(Cambridge Austerdalsbreen Expedition)

J.W. GLEN

Physics Department, Birmingham University

SUMMARY

Measurements made on the snout of Austerdalsbreen, Norway, show that the glacier is undergoing compressional strain even in the last 20 m. of its length. The ice velocity at the actual ice edge is about 10 m. per year, and due to compressional strain-rate this increases to 30 m. per year at a distance 220 m. up-glacier. Although it retreated 37 m. between 1958 and 1959, the glacier was moving faster in 1959 than it had been in 1958. The longitudinal strain-rate varies markedly over the region; the only correlation that can be found is between it and the vertical velocity of the ice, a result which suggests that the strain-rate variations are of the same sign throughout the thickness. A measurement of the surface strain-rate tensor using a square of stakes did not give a very accurate result owing to the marked variations of strain-rate with distance, even within the 18 m. dimensions of the square. The largest principal strain-rate was compressive and of magnitude 0.19 ± 0.02 per year, corresponding to a maximum principal stress of 0.81 ± 0.07 bars.

1. INTRODUCTION

The snout of a valley glacier is a region of great glaciological interest. The geomorphologist is interested in its movement in connexion with the formation of end moraines, and the glacier theorist is also concerned to predict its deformation and its fluctuations as a result of climatic change. Lliboutry⁽¹⁾ has recently discussed possible forms that the motion of a glacier snout could have on the basis of the perfectly plastic model. A glacier could have a relatively dead snout which remains stationary while the ice is compressed behind it, or the snout could be moving forward as a rigid body, all moving at the same velocity, so that the *deformation* of the ice, as opposed to its movement, may occur only further up-glacier. The third possibility is that the ice is both moving and deforming right down to its end.

There have not been many measurements of the variations in ice velocity close to a glacier snout, and so measurements of this kind were included in the programme of the Cambridge Austerdalsbreen Expedition in the summers of 1958 and 1959. Austerdalsbreen is a valley glacier formed by two ice falls, Odinsbreen and Thorsbreen, which flow from the ice cap known as Jostedalsbreen in western Norway. Measurements of its velocity and deformation have been made just below Odinsbreen ice fall⁽²⁾ and also, more recently, in a region in the middle of the glacier, and less accurate measurements of longitudinal strain have been made all down the length of the glacier. Further its retreat is known from measurements made annually under the direction of the Norsk Polarinstitutt; it is retreating rapidly and has been doing so for many years now⁽³⁾. The snout can thus be taken as typical of a retreating glacier, a situation which might lead us to expect relatively stagnant conditions at the snout.

The snout itself is at present in a place where the glacier is curving to the right, being just above a rock bar across the valley. In between July 1958 and July 1959 the glacier had retreated a distance of 37 m., leaving an area of morainic debris between itself and the smooth bedrock of the rock bar which it was just reaching in 1958.

2. MEASUREMENTS MADE

In July 1958 measurements were made of a line of stakes running along what was estimated to be approximately the line of flow and ending just to one side of the stream emerging from under the glacier. This line was aligned onto one corner of a very large boulder beside the stream in the valley below (painted with the letters BB); the spacing of the stakes was 15 m at the lower end, extending to 29 m. further up-glacier, through, owing to the crevassed nature of the surface, the distances were not all uniform. The total length of the line, measured along the glacier surface, was 219 m. and the bottom stake was about 5 m from the ice edge. The relative movement of the stakes was measured by taping and levelling using techniques similar to those employed in earlier surveys at the foot of the ice fall (²), and the movement of the stakes was also measured by theodolite observations from three stations on the rock bar. After analysis of these observations, it was found that the accuracy of the velocity determined from theodolite measurements was not as great as that from the taping and levelling; this is probably due to two reasons, first the rays used were frequently nearly tangential to the rock and ice surfaces, so that the great variations in temperature occurring on a glacier in summer can lead to serious refraction errors, and secondly it was noticed that the longer the delay between the theodolite observations from the different stations, the worse was the error in the position of the stakes, and this effect was not eliminated by allowing for a steady flow of the glacier, so that it may well be that the movement of the glacier is not steady enough to allow interpolation of the changes in theodolite angles. If the movement of a glacier is jerky, then theodolite observations taken from different stations at different times will inevitably give poor results for the glacier flow.

However the theodolite observations do give one piece of information that is not obtainable from the taping and levelling; they yield the direction of movement of the stakes, whereas the taping and levelling only tell us the components of velocity in the vertical plane through the line of the stakes.

Other measurements made in 1958 included measurements of stakes placed either side of bands which looked as though they might be active thrust planes, and measurement of movement close to the ice edge by means of a stake projecting right through the ice into the gap between the ice and the bed.

In 1959 a search was made for the stakes left behind in 1958, which included two stakes bored in to a depth of 6 m in the hope that they would survive to give measurements of velocity and strain over a full year, but unfortunately, owing to heavy ablation and relatively low snowfall in the intervening months, no stakes were still in position in the ice. Indeed all except one appeared have rolled far from their original sites. That one was the top stake of the 1958 line (stake 870), which was found in 1959 in a small hollow recognisably similar to one near it in 1958; this stake alone gives some indication of the flow in a full year. The main measurements made in the snout area in 1959 were the taping and levelling of a square of stakes inserted into the middle of the region that had in 1958 been used for the line of stakes. The object of these measurements was to determine the strain-rate tensor at the glacier surface, and the technique used was the same as had been employed for this purpose at the foot of the ice fall (⁴). In addition a few stakes were placed along the line for velocity measurements. Both in 1958 and 1959 the rate of ablation was measured at all the flow stakes.

3. RESULTS

Although the taping measurements only covered a period of 12 days both movement and strain were quite sufficient to be measured throughout the length

of the 1958 line of stakes. Unfortunately, owing to bad weather, the levelling observations cover an even shorter period; 7 days for the lower stakes and only 4 days for the upper ones, but here too the changes appear reasonably significant. Table I shows results for the horizontal component of velocity parallel to the line of stakes and also for the longitudinal strain-rate measured parallel to the glacier surface.

TABLE 1

Velocity and strain of stakes on snout of Austerdalsbreen. Bottom stake, 860, was 5m from ice edge.

Stake	Distance from Stake 860 m	Height above Stake 860 m	Horizontal component of velocity m/yr	Longitudinal strain-rate per year
860	0	0	10.1	— 0.06
861	16	3.6	11.0	+ 0.01
862	31	6.9	11.2	— 0.13
863	46	11.1	12.9	— 0.10
864	61	16.0	14.1	— 0.06
865	90	24.9	16.0	— 0.16
866	109	29.5	18.9	— 0.08
867	131	34.8	21.0	— 0.14
868	161	44.4	25.2	— 0.10
869	190	54.7	28.0	— 0.09
870	219	61.1	30.8	

It will be seen that, apart from one leg, all the strains were compressive, and that even with that leg the effect of the vertical velocity (determined from the levelling) was such as to make the horizontal velocity of the ice increase monotonically up the glacier. This increase by a factor three in the last 220 m of the glacier is quite striking, and shows the large amount of deformation the ice is undergoing. Even in the last stake interval on the ice, i.e. over the distance 3 to 19 m from the actual end of the ice, there was still a strain rate of 0.064 per year compressive. This is a remarkable result, as at its end the ice is not touching the bed, there being a gap a few centimetres wide over almost all the edge. While such a gap persists under the ice it seems inconceivable that the ice should be compressing itself longitudinally, but right down to the place where it loses contact with the bed it would seem that such strain is taking place; the ice of the snout is not dead, nor is it being pushed forward as a rigid mass for any distance more than a few metres from the ice edge.

The results of the measurements taken in 1959 are shown in Table 2. The location of the square was such that stakes 964, 960 and 962 occupied positions in space fairly close to those occupied a year before by stakes 864, 865 and 866, though the ice surface was about 3 m lower at stake 964 and about 7 m lower at the other two stakes than it had been the previous year. The other two stakes of the square were stakes 961 and 963, which were to either side of stake 960. It will be seen that the strain-rates in 1959, though of the same order as in 1958, are not very similar, that for the top interval being about double, and that for the lower interval about half, the values of the previous year.

TABLE 2

Strain-rates measured on the square of stakes.

Stake Interval	Strain-Rate per year	Stake Interval	Strain-Rate per year	Stake Interval	Strain-Rate per year	Stake Interval	Strain-Rate per year
960-961	0.09	961-962	- 0.04	960-962	- 0.32	962-963	- 0.23
960-963	- 0.02	963-964	- 0.05	960-964	- 0.04	964-961	- 0.02

Apart from these measurements of surface strain, the most interesting results obtained in 1959 were of the velocity of flow. The stake 870 had moved a distance of 22 m over the year—or more precisely 353 days—between the last 1958 survey and when it was found in 1959. Subsequently it was re-erected and theodolite observations showed it to be moving at 38 m/year, so that this stake must have slowed down during the winter, but reaccelerated to a velocity greater than that it had had in the preceding summer. Theodolite observations on the stakes of the square and also on one further stake between the square and the end of the glacier showed that these stakes were also moving faster than the ice in the same position had been a year previously—and this despite the fact that they were of course nearer to the end of the glacier as a result of the retreat. The results for their velocities, together with the velocities of the stakes most near to their position in space in the preceding year, are shown in Table 3

TABLE 3

Velocity of stakes in 1959 compared with velocities of stakes nearest to their positions in 1958.

Stake	Distance from end of glacier m	Height above reference mark m	Velocity m/yr	Stake nearest in 1958	Velocity m/yr
965	18	6.4	13	863	13
964	40	13.7	17	864	14
960	58	19.0	20	865	16
962	76	23.0	25	866	19
870	163	51.8	38	869	28

If the comparison had been made with stakes the same distance from the end of the glacier, the acceleration would have been even more marked. The conclusion to be drawn from this is that in 1959 the snout area of Austerdalsbreen was moving faster than in 1958.

4. ANALYSIS OF RESULTS

The measurements made on the line of stakes cannot be analysed in the same way as were the measurements on the line at the foot of the ice fall (²); there are several differences between the conditions at the snout and at the foot of the ice fall which complicate the situation and render invalid the assumptions made by Nye in that analysis. First, in the derivation of the formula from which he calculates the strain-rate, Nye assumes that various quantities do not vary rapidly with distance down the glacier; near the snout the depth of the glacier and the curvature of its bed are varying rapidly in distances of the order of the depth. Secondly, in deriving his general formula, Nye assumes that the shear stress on the bed is approximately constant; near the snout this assumption is going to break down as the ice loses contact with the bed, for the shear stress on the bed must fall to zero as this contact is completely lost. Finally, Nye's analysis assumes that the glacier is in a steady state, and this assumption is probably not too bad if the glacier is an annually repeating state. However near the snout the state is very far from being annually repeating, since in the year in question the rate of retreat of the glacier over the year was greater than the flow velocity of the snout ice. For all these reasons it is not to be expected that such an analysis will be successful; if it is attempted it gives values for the strain rate an order of magnitude greater than those observed and this can be interpreted as meaning that the glacier in this snout region would have to compress itself much faster than it is if it were to compensate adequately for the present amounts of ablation.

In the absence of theoretical predictions concerning the snout measurements the best analysis is probably an empirical search for correlations between the various parameters such as ablation, surface slope, vertical velocity and longitudinal strain-rate. Only two of these show any marked correlation, vertical velocity and longitudinal strain-rate. The vertical velocity seems to be a maximum when the longitudinal strain rate is most compressive (See Fig. 1). This seems a reasonable correlation, since a compressive strain rate, if continued throughout the glacier thickness, should be making the ice thicker and hence give the ice a higher vertical velocity at the surface.

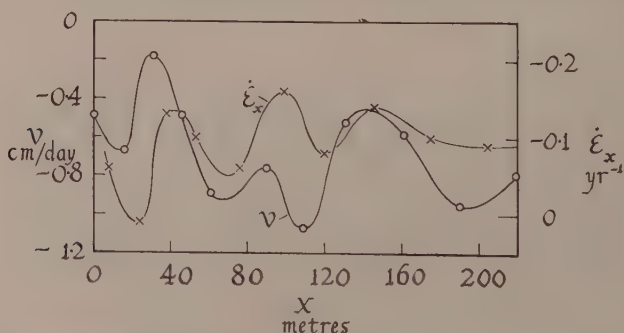


Fig. 1 — Variation of vertical velocity v and of longitudinal strain-rate $\dot{\epsilon}_x$ as a function of the distance x up the glacier from the lowest stake. The strain-rates have been plotted with negative (compressive) strain-rates upwards so that the correlation between vertical velocity and compressive strain-rate can be best assessed.

Perhaps this correlation can be taken as indicating that the fluctuations of strain rate in the snout, unlike those below the ice fall, represent variations of the same sign throughout the glacier depth and are not due to bending.

The square of stakes measured in 1959 can be used to determine the strain-rate tensor at the surface of the glacier using the method described by Nye (4). When this is done, it is found that the strain varies markedly within the square, and that consequently the accuracy of the resulting strain components is not as great as in the earlier determinations. Using the same notation as that paper, the strain components are:

$$\dot{\epsilon}_x = -0.18 \quad \dot{\epsilon}_{xz} = 0.04 \quad \dot{\epsilon}_z = 0.02 \text{ per year}$$

with a standard error of about ± 0.02 per year. If this is analysed further to give the principal strain-rates, the result obtained is:

$$\dot{\epsilon}_1 = -0.19 \quad \dot{\epsilon}_2 = 0.16 \quad \dot{\epsilon}_3 = 0.03 \text{ per year}$$

with the same accuracy as for the components above, and the angle φ that the axis Oz makes with $\dot{\epsilon}_3$ is $-11 \pm 5^\circ$. Finally if we use these strain rates to compute the principal stresses using the laboratory flow law, the results are:

$$\sigma_1 = 0.81 \pm 0.07 \quad \sigma_3 = -0.30 \pm 0.07 \text{ bar.}$$

The relatively large inaccuracy in all these results reflects the fact that in the snout area the conditions are changing so rapidly that even within the small square size used (18 m sides), conditions are far from uniform. A smaller square could not be used without reducing the changes of length to very small amounts.

The observations that show the glacier to be moving faster in 1959 than in 1958 despite its decreased depth and shorter length are further indication that the movement of a glacier snout depends on conditions upstream at earlier times and not only on the local position of the snout. Nye (5) has recently developed a theory of the response of glaciers to changes in conditions that shows how waves of increased flow can be propagated down a glacier and will give very complex effects at the snout; it would seem that the snout of Austerdalsbreen is at present receiving a surge of increased movement, but whether this will be sufficient to stop retreat seems doubtful; for that a very large increase in flow velocity would be needed.

5. ACKNOWLEDGEMENTS

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SUR UNE NOUVELLE METHODE D'INVESTIGATION EN GLACIOLOGIE (1)

R. MILLECAMPS (France)

RÉSUMÉ

Des mesures de photogrammétrie terrestre effectuées à cadence accélérée ont montré en 1954-56 que l'on peut suivre de 12 en 12 heures le micromouvement et les déformations superficielles d'un tronçon de glacier. L'auteur a cherché un moyen de réaliser des investigations analogues sur l'épaisseur du glacier sans créer de perturbation physique, thermodynamique ou mécanique dans le milieu. Il expose le principe et les tout premiers résultats des expériences mises en chantier à la Mer de Glace en 1957 et poursuivies depuis avec succès.

SUMMARY

Terrestrial photogrammetry measurements made at the rate of one pair of pictures every 6 hours have shown, in 1954-56, that it is possible to observe, every 12 hours, the micromovement and the surface deformations of a glacier section. The author has sought to realize same measurements across the depth of the glacier without any physical, thermodynamical or mechanical perturbation in the material. He shows the principle and the very first results of experiments carried out in the Mer de Glace glacier during 1957 and since successfully continued.

L'étude des mouvements et de l'écoulement des glaciers repose essentiellement sur des observations de terrain. A l'aide de dispositifs appropriés on réalise un balisage du secteur (ou de l'ensemble) étudié, et l'on suit les évolutions de ce balisage dans le temps et dans l'espace. On cherche ensuite à interpréter et à définir mathématiquement le phénomène enregistré.

La Photogrammétrie terrestre et la photogrammétrie aérienne rendent en particulier les plus grands service en permettant d'effectuer d'une manière élégante les mesures et la comparaison des résultats à des intervalles de temps divers. Ces techniques présentent l'avantage de laisser au glaciologue des documents que l'on conserve et que l'on peut à loisir extraire des archives pour en reprendre ou en compléter l'étude.

S'inspirant de ces idées générales, l'auteur a voulu suivre *dans le détail* les évolutions d'un tronçon du plus grand glacier français, la Mer de Glace (massif du Mont Blanc). Entre 1954 et 1959 il a été successivement procédé à deux types distincts d'investigations.

Les premières années, on a suivi l'écoulement en surface d'un tronçon du glacier, long environ de 1 kilomètre (choisi pour des raisons de commodités et d'accès, dans la partie aval), en mettant à profit les avantages de la photogrammétrie terrestre; on a procédé, alors, à des prises de vues effectuées à *cadence accélérée* : un couple de clichés, pris simultanément, toutes les 6 heures (pendant 52 jours en 1954, par exemple).

Les résultats obtenus ont montré que les déplacements enregistrés en 12 heures pour chacun des repères utilisés sont supérieurs aux erreurs d'expérience, et qu'il est aussi possible de suivre par le détail l'écoulement des différentes zones de glace dont l'ensemble constitue le tronçon expérimenté. Dès lors, l'auteur a recherché des expériences analogues mais qui rendaient compte, cette fois, des phénomènes internes et non plus uniquement superficiels : mesure de l'écoulement et des déformations de la glace au sein même du glacier.

(1) Cette communication, prête depuis Novembre 1959, n'a pu, pour des raisons indépendantes de ma volonté, être présentée au Symposium Antarctique de Buenos-Aires (S.C.A.R., 16-25 Nov. 59).

De nombreuses études ont été menées dans cette voie. Nous citerons essentiellement, pour en résumer l'esprit, les expériences de Perutz, Gerrard et A. Roch ⁽²⁾ de 1948 à 1950 au Jungfraufirn (Oberland bernois, Suisse) et les remarquables études de déformation d'une galerie au Jungfraujoch conduites par le Professeur Haefeli ⁽³⁾ de Zürich, et son équipe.

Il convient cependant de remarquer qu'il s'agit d'expériences qui portent sur du névé, c'est-à-dire sur un milieu et un matériau différents de la glace du dissipateur; que les déformations du tube d'acier de Perutz, enfoncé jusqu'à 137 mètres dans du névé, ne traduisent peut-être pas toutes celles de ce névé (qui peut par exemple «fluer» autour de ce tube, lequel résiste par lui-même aux contraintes qu'il subit); enfin, que l'existence même d'une galerie au sein de la masse que l'on étudie modifie dans une certaine mesure les divers paramètres (physiques, mécaniques, etc...) de cette masse et, par conséquent, le comportement de celle-ci.

L'auteur a donc cherché à mettre au point un type d'observation et une méthode de mesures qui, pour compliquées qu'elles paraissent être, et valables seulement pour l'instant dans la glace de glacier tempéré ⁽⁴⁾, rendraient nulles ou négligeables les perturbations du genre de celles qui viennent d'être évoquées. Le but de cette entreprise étant de saisir, de suivre et de mesurer à volonté les modalités de l'écoulement de la glace sur la plus grande partie de l'épaisseur d'une langue glaciaire, on a délimité, en 1957, un échantillon prismatique de glace de 13.000 m³ sur lequel ont lieu, exclusivement, toutes les opérations.

L'expérience se déroule depuis la fin de l'été 1957. Le secteur choisi, pratiquement exempt de crevasses ou fissures importantes, se situe à environ 2 km en amont du front du glacier il correspond à une portion de glacier pratiquement rectiligne dont le socle ne paraît pas (d'après les sondages effectués par E.d.F.) présenter d'anomalies et dont l'écoulement semble être, à priori, plus simple et plus classique que dans les autres tronçons du glacier. La largeur d'une rive à l'autre est de 800 m environ, la surface médiane du tronçon, exempte de moraine, étant de l'ordre de 400 m. C'est au milieu de cette zone médiane de glace vive et aussi exactement que possible dans l'axe du glacier que le prisme expérimental de glace a été défini.

Dans ce but on a délimité à la surface du glacier un carré ABCD de 10 m de côté, tel que les côtés AB et CD soient alignés sur la direction générale d'écoulement, et que le côté CD soit pratiquement sur l'axe du glacier à cet endroit. En utilisant chacun des 4 sommets du carré comme point de départ, on a effectué 4 forages (aussi verticaux et parallèles que possible) de 8 cm de diamètre sur 140 m de profondeur. Ce prisme de glace étudié est défini à l'origine par ces 4 trous qui matérialisent en quelque sorte ses 4 génératrices principales. Dans chacun de ces forages et à partir de la surface libre du glacier, considérée (à des corrections près) comme niveau de référence, on a descendu et installé des céramiques piézo-électriques émettrices et réceptrices d'ultrasons, à raison d'une céramique tous les 10 mètres. On compte donc théoriquement 4 céramiques par niveau de 10 mètres. Le trou C présente la particularité d'être équipé de 0 à 60 m, d'une céramique tous les 5 mètres de profondeur.

Ces céramiques au titanate de baryum sont de très petites dimensions : 40 × 40 × 7 mm et constituent pratiquement des *points* au sens mathématique du mot par rapport aux dimensions du prisme de glace et, à *fortiori*, par rapport à celles du glacier. Elles sont munies d'un système d'ancrage automatique dans la glace qui est déclenché depuis la surface, au cours de l'installation, au moment où la profondeur et l'orientation désirées sont obtenues; elles ne peuvent plus, dès lors, avoir de mouvement propre de rotation ou de translation par rapport au volume de glace qui entoure chacune d'elles. Elles sont connectées à volonté à l'appareillage électronique installé

(4) Des expériences sont actuellement en cours dans du névé et de la glace froide non mouillée.

en surface par un jeu de câbles coaxiaux. Cet appareillage important est convenablement isolé et installé sur un plancher à l'intérieur d'une grande tente implantée sur le glacier; il convient de noter que les câbles peuvent éventuellement s'allonger d'au moins 50% au cas où une crevasse se formerait brutalement dans ce secteur et que leur revêtement externe en matière plastique spéciale empêche toute prise de la glace sur le câble qui coulisse à la moindre traction. Ils ne sont, en fait, jamais tendus. Il faut également remarquer que les forages n'ont pas été exécutés sur toute l'épaisseur du glacier (profondeur de 140 m sur une épaisseur totale de 175 m), qu'ils se remplissent d'eau automatiquement au fur et à mesure que le forage est effectué, que cette eau ne regèle que partiellement et en surface seulement au cours des hivers qui suivent, et, enfin, que le diamètre de ces trous diminue lentement en fonction du temps, la glace enserrant progressivement de plus en plus les céramiques. L'eau qui existe en permanence au sein de la masse de glace, quelle que soit la saison, (glacier de type tempéré) assure un lien acoustique excellent entre les faces émettrices et réceptrices des céramiques et le milieu qui les enrobe. De plus, les signaux se propageant en travers du prisme, on peut admettre que la glace qui constitue ce dernier n'a subi aucune perturbation ni de la part des forages ni de la part des céramiques.

L'expérience consiste à mesurer avec précision les temps de transit d'un signal ultra-acoustique allant d'une céramique prise comme émetteur à plusieurs autres prises comme récepteurs et ceci à une cadence régulière dans le temps (toutes les 24, 12, 6 heures ou même toutes les heures). Ce temps de propagation permet de connaître les distances qui séparent les céramiques et les variations de ces distances, ceci sans perturbation du milieu. La vitesse des signaux qui est connue au départ avec une certaine approximation et que nous persions, dès 1957, légèrement variable avec la profondeur, peut être dans le calcul soit éliminée par des mesures relatives, soit redéterminée expérimentalement et mathématiquement en utilisant la propriété géométrique suivante des points de l'espace : si les distances a, b, c , d'un point 0 à 3 points connus et non alignés de l'espace A, B, C , sont connues, la position du point 0 est connue à une symétrie près (incertitude qu'on élimine facilement dans notre cas). Si la distance du point 0 à un 4^{me} point connu D non situé dans le plan ABC est mesurée et ensuite calculée, les deux résultats doivent concorder; sinon la valeur-paramètre qui a servi à calculer les distances est à reconsidérer. Dans notre cas, ce paramètre est la vitesse de propagation des signaux dont la valeur est redéfinie pour chaque étage de 10 m.

De plus, l'expérience peut encore être réalisée lorsque aucune distance n'est connue avec précision, pourvu que l'ensemble soit connu approximativement.

Prenons pour référence un trièdre $Oxyz$ ayant son origine placée sur une céramique. Soit :

- 1, 2, 3, ... les numéros des céramiques voisines,
- $(x_1 \ y_1 \ z_1), (x_2 \ y_2 \ z_2), \dots$ les coordonnées de ces céramiques,
- $d_1, d_2, d_3 \dots$ les distances à l'origine des céramiques,
- t_1, t_2, t_3, \dots les temps mesurés correspondants,
- d_{12} , et t_{12} les distances et temps entre 1 et 2, etc ...
- c la vitesse de propagation au niveau de 0.

On a :

$$x_1^2 + y_1^2 + z_1^2 - (ct_1)^2 = 0$$

$$x_2^2 + y_2^2 + z_2^2 - (ct_2)^2 = 0$$

$$(x_2 - x_1)^2 + (y_2 - y_1)^2 + (z_2 - z_1)^2 - (ct_{12})^2 = 0$$

soit 3 équations, 7 inconnues, 2 distances de 0, 1 distance entre les céramiques 1 et 2.

D'une façon générale, on peut montrer que si on mesure n distances à partir de 0, on obtient $3n + 1$ inconnues et $1 + 2 + 3 \dots + n$ équations, soit un rapport (nombre d'équations)/(nombre d'inconnues) qui croît avec n jusqu'à dépasser l'unité. Par exemple, pour $n = 6$, on a 21 équations et 19 inconnues. Lorsqu'on obtient plusieurs solutions pour une inconnue, la connaissance approximative que nous en avons toujours permet de lever l'incertitude.

Expérimentalement, on mesure des temps de transit relativement courts si l'onde s'est propagée dans la glace, beaucoup plus longs (sensiblement 2,5 fois) pour les ondes qui se sont propagées dans l'eau qui remplit un trou au début de l'expérience. Dans un cas particulier, nous avons trouvé à 60 mètres de profondeur des temps respectifs de 3.293 et 7.741 ± 3 microsecondes.

Ces complications sont un inconvénient de la méthode mais il n'est pas possible, à notre connaissance, d'utiliser des signaux autres qu'acoustiques. Les ondes hertziennes en particulier ne donneraient que des résultats fantaisistes la glace étant une sorte de semi-conducteur avec plus ou moins d'impuretés. Incidemment, en effet, nous avons pu constater des variations aléatoires et sans rapport direct avec les distances, allant jusqu'à 20 microsecondes dans le temps de propagation des ondes hertziennes d'une céramique à l'autre.

1. PRINCIPE DE L'APPAREILLAGE ÉLECTRONIQUE

Les signaux émis et reçus à 65 KHz sont enregistrés au moyen d'une caméra 35 m/m à déroulement rapide et continu (4 à 12 m/sec) que nous avons construite nous-mêmes. Cette caméra filme un ensemble de 11 spots très lumineux de 11 tubes cathodiques alimentés sous 6.000 volts dont on supprime temporairement le balayage horizontal, remplacé par le déroulement du film. L'utilisation de chronomètres électroniques classiques (qui aurait évité la lourdeur du dépouillement) s'était révélée, en effet, impossible, après les essais de 1957 pour diverses raisons, dont les échos, les caractéristiques des céramiques, la forme des signaux sismiques éventuellement utilisés, le niveau et la forme des parasites acoustiques ou telluriques. De plus, et beaucoup plus que de simples chiffres, les enregistrements sont riches d'informations toujours disponibles, sur le comportement de la glace comme sur le fonctionnement même de l'appareillage, et ceci constitue un «auto-contrôle».

Le cerveau de l'appareillage est constitué par une «échelle électronique» qui fournit les signaux temps et les commandes nécessaires. Le temps est donné par un oscillateur à quartz thermostaté à 60°C, dont la stabilité relative est au moins supérieure à 10^{-5} . Après multiplication convenable les périodes fournies par le quartz sont transformées en tops-repères des divers intervalles de temps utilisés (10, 80, 640 et 10.240 microsecondes) qui apparaissent sur chaque écran en superposition avec les signaux étudiés. Chaque top 10.240 commande l'émission d'impulsions ultra-acoustiques par un émetteur approprié de 1 KW qui commande lui-même le fonctionnement de l'armoire aux onze tubes cathodiques. Après de nouvelles mises au point effectuées au début de l'année 58, le temps de lancement de l'impulsion est connu à moins de une microseconde près. Pour la préparation de chaque mesure par un examen visuel direct des tubes cathodiques, l'échelle fournit un balayage approprié commandé par les tops-temps ci-dessus. De plus, il est possible de sélectionner telle partie plus intéressante de l'image : on «efface» alors le reste de l'image et on dilate la bande considérée. On peut aller ainsi à une vitesse d'examen de 15 m/m pour 10 microsecondes (soit 1,5 cm d'écran pour 3 cm environ dans le glacier).

Les 9 récepteurs sont basés sur le même principe que celui des essais de Mai 1957. Cependant, des précautions spéciales ont été prises pour compenser les diverses longueurs possibles de câbles et pour réduire l'importance du temps de retard des

céramiques. On peut montrer, en effet, que la constante de temps θ d'un circuit accordé électro-acoustique du type piézo-électrique est donnée par la relation :

$$\theta = \frac{2Q}{\omega} \text{ où } Q \text{ est le facteur de surtension et } \omega \text{ la pulsation du signal, } \omega \text{ étant}$$

imposé par les conditions de propagation, il n'était possible que de réduire le facteur Q réel ou apparent.

2. RÉSULTATS

Il a été ramené des campagnes de 1957 et 1958, 632 films d'enregistrement qui représentent environ 15.000 mesures; le dépouillement, suivi de très nombreux calculs de résolution des systèmes d'équations cités, en est extrêmement long et délicat; il est toujours en cours, mais une partie des résultats est en voie de publication.

Début 1958, c'est-à-dire 2 mois après la fin de la dernière campagne de mesures sur le glacier, il était déjà possible de donner les premières conclusions suivantes :

1° Il apparaît des phénomènes annexes, à savoir :

a) des signaux parasites à des niveaux et avec des formes variables en fonction de l'heure et même de la profondeur dans le glacier. Ils sont plus importants entre 1 et 4 heures du matin. Ils paraissent dus à une activité ionique et électronique d'origine intérieure ou extérieure au glacier;

b) il se superpose localement et d'une manière aléatoire des signaux d'amplitude relativement grande et de forme caractéristique des bruits acoustiques.

2° Au point de vue de la mesure des distances entre céramiques et compte tenu de ces phénomènes,

a) la précision atteinte est maintenant de 1 cm, soit une précision relative de 1‰;

b) les variations des distances avec le temps semblent liées à la direction et dépendent de la profondeur;

c) la vitesse de propagation des signaux est une fonction de l'épaisseur de glace traversée.

Il semble bien que la vitesse de propagation des ultrasons doive augmenter avec la profondeur atteinte dans le glacier (qui est de type tempéré). Mais celui-ci étant dans toute sa masse à température de fusion, est près de zéro degré. Des expériences russes⁽⁵⁾ de haute précision effectuées au laboratoire à des températures très voisines du point de fusion, ont montré que la vitesse décroît d'une façon importante, passant de 3.500 à 1.700 m/s lorsque la température tend vers zéro degré, puis remonte brutalement à plus de 3.000 m/s pour une température de quelques centièmes de degré au au-dessus de zéro. Il y aurait donc lieu de penser qu'à des profondeurs égales ou supérieures à 70 mètres le milieu est constitué par un mélange de 2 phases, l'une solide, l'autre liquide, la seconde, de par son importance, étant à l'origine de la diminution de la vitesse de propagation des signaux que nous notons effectivement après de multiples mesures effectuées sur nos enregistrements;

d) l'ensemble des premières lectures effectuées en 1958 et qui pouvaient être comparées aussitôt à celles de 1957, a montré une diminution générale de toutes les longueurs de parcours des signaux entre céramiques, ou, plus exactement, une diminution générale de tous les temps de transit : ce fait indique indiscutablement qu'un polyèdre défini géométriquement par 8 céramiques disposées 4 par 4 à 2 niveaux différents, a subi une contraction. On peut en déduire, soit que la glace contenue dans ce volume géométrique s'est tassée sur elle-même, soit qu'elle s'est écoulée à travers ce volume en provoquant des changements de position des céramiques, soit que, avec le temps, le milieu ait subi un changement de structure qui aurait eu pour conséquence une augmentation appréciable de la vitesse de propagation des signaux.

Il est actuellement difficile de se prononcer déjà sur cette question.

Mais en ce qui concerne l'allure elle-même des déformations subies par chacun des polyèdres définis par 8 céramiques le dépouillement actuellement exécuté montre qu'il est possible de mettre en évidence les «petits mouvements et déformations» d'un volume de glace «in situ» en fonction du temps et, en particulier, dans un intervalle de quelques jours seulement où l'on suppose alors qu'il n'y a pas de changement de structure de milieu. Dans ces conditions on a pu constater, entre autre, des débuts de rotation d'un volume de glace situé, par exemple, à une vingtaine de mètres de profondeur, autour de 2 axes particuliers.

L'exploitation de cette longue expérience fait ressortir maintenant que, pour la première fois, sur la base de mesures physiques de précision, nous pouvons apporter sur un glacier des données objectives. Les résultats qui commencent à sortir vont donc confirmer ou infirmer les vues mathématiques des glaciologues théoriciens.

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A PROPOS DES «VAGUES» DES GLACIERS : ASPECTS OBSERVES SUR LA LANGUE ET SUR LE FRONT DU GLACIER DU LYS (MONT ROSE)

G. ALIVERTI

Istituto Universitario Navale
Napoli Gabinetto di Meteorologia e Oceanografia

SUMMARY

The author reports on some morphological peculiarities of the Lys Glacier (M. Rosa) observed during several visits to the glacier tongue in the course of the last teens. Such peculiarities are to be ascribed to the presence of the ice-fall above the tongue and, hence, to be defined as « wave ogives » in the meaning described by F. Nye.

Plusieurs travaux, parus récemment, ont développé des considérations et des discussions sur les phénomènes dits des «vagues des glaciers» (wave ogives), vagues se vérifiant sous les «chutes» de glace du bassin de dissipation, c'est-à-dire dans la langue, au pied d'une portion de glacier fortement inclinée.

J.F. Nye ⁽¹⁾ expose dans le volume «Symposium de Chamonix» (1958) une théorie assez convaincante sur la formation de ces vagues; et dans le fascicule n° 25 du J. of Glac. (1959) ⁽²⁾ il revient sur ce sujet en démontrant, sur la base de mesures prises sur un glacier, que la formation des vagues peut être expliquée par l'action combinée de la déformation plastique et du mécanisme de l'ablation. Ces travaux et d'autres encore m'ont rappelé les différents aspects que j'ai observés au cours des dernières années, pendant plusieurs visites faites en été, depuis 1941 jusqu'à maintenant, au Glacier du Lys dans le groupe du Mont Rose.

Le bassin de dissipation de ce glacier commence à une cote de 3250 m environ. Il est exposé, ainsi que tout le glacier, au sud et, de ce fait, il est soumis pendant l'été à des phénomènes intenses d'ablation. La zone du glacier dénommée «plateau» se trouve à une altitude d'environ 2500 m et elle est formée par une étendue presque plane située sous la partie très inclinée du glacier qui de 3250 m de cote arrive aux 2500 m du plateau. La portion de langue au-dessous de la chute, encore longue et imposante au cours des années de 1925 à 1935-40, est aujourd'hui très raccourcie et s'étend peu au-dessous du plateau; en effet, le front se trouve maintenant franchement au-dessus du Roccione de Salzen, c'est-à-dire au-dessus de 2300 m d'altitude (v. fig. 1). Elle s'est par conséquent réduite à peu près à deux cents mètres seulement et, par rapport à 1921, elle s'est raccourcie au total de 500 m environ. Le volume de glace qui forme la langue est aussi, naturellement, réduit et l'on peut estimer qu'une épaisseur de glace de plusieurs dizaines de mètres ait disparu.

Mes visites au Glacier du Lys, à fin d'étude, commencèrent en 1941 pour observer la structure de la glace. Au cours des années 1941 et 1942 je réunis plusieurs observations sur le névé, derrière la Capanna Gnifetti, et sur la langue dans la région du plateau; en 1947 je retournai sur le glacier en vue de recueillir des observations sur la zone de séparation entre le névé et la langue à la cote de 3250 environs et, grâce à elles, je pus mettre en évidence l'influence décisive du climat local et des facteurs météorologiques sur cette partie du glacier. En Août 1950 et 1951 je visitai le plateau pour y chercher les *cônes de glace* et j'observai les phénomènes d'ablation et de mouvement, documentant photographiquement mes observations et en particulier la rotation de 90° d'une *petite colline de glace*, rotation qui se produisit dans le bref intervalle de neuf jours. Ce phénomène de mutation si rapide induisit le prof. L. Solaini

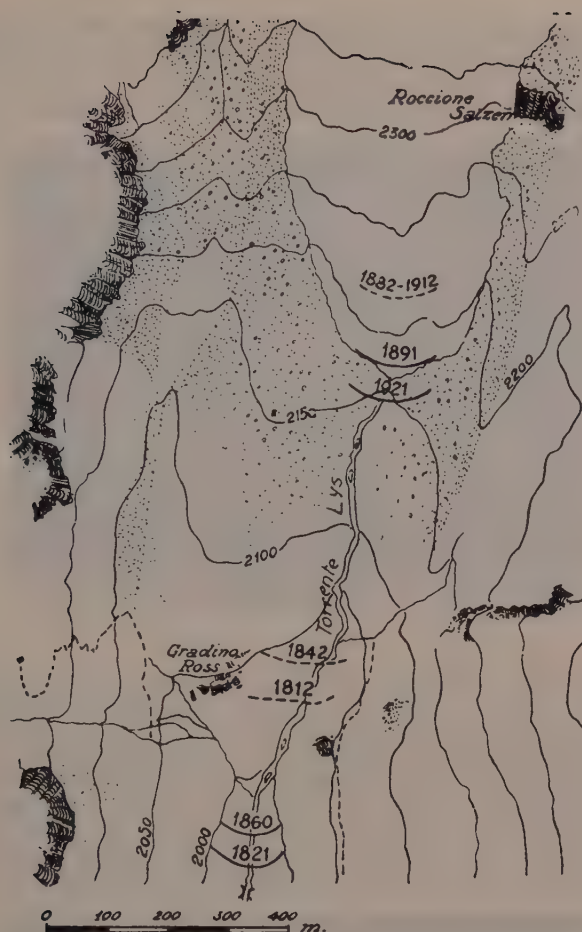


Fig. 1 (Selon U. Monterin, Boll. Comitato Glaciologico Italiano, 1932)

du Polytechnique de Milan à effectuer un levé photogrammétrique du plateau du Lys, ce qui eut lieu au mois d'Août 1953. Les conditions climatiques peu favorables à l'ablation ne permirent pas d'observer, d'après deux levés faits à neuf jours de distance, des phénomènes aussi imposants que celui photographié par moi en Août 1951, mais elles mirent en évidence dans certaines positions des variations sensibles des lignes de niveau.

Sur toutes ces observations (1941-1953) je présentai un bref rapport à la réunion de l'Association Internationale d'Hydrologie Scientifique qui eut lieu à Rome en 1954.

En Août 1955 je retournai sur le plateau et, de cette visite, je ne fis aucune relation; le glacier paraissait plus encore consommé et en mauvais état que les années précédentes et son front laissait découvert le Roccione de Salzen. J'avais, toutefois, observé un fait qui aurait pu faire soupçonner une faible croissance du glacier, le front semblait

assez gonflé et surtout il était en train de former une *moraine frontale*. La fig. 2 représente une partie de cette moraine. Mais, au cours des années suivantes, la langue a continué de diminuer dans ses dimensions et à se rétrécir, et, au mois d'Août 1959, je pus constater que le front était sensiblement au-dessus du Roccione de Salzen.



Fig. 2

La formation de la moraine frontale est un des faits auxquels je faisais allusion au commencement de cette note et que j'estime une conséquence des «vagues» du plateau. Et qu'il en soit ainsi, c'est-à-dire qu'il existe même sur la langue du Glacier du Lys la formation d'ondulations produites par la chute de glace située au-dessus du plateau, est une déduction que l'on peut tirer d'uatres observations faites au cours des dernières années. Dans une des photographies prises par le prof. Solaini en 1953 en regardant le glacier de la moraine de droite, on peut voir clairement deux ondulations longitudinales de la première nappe de langue recouverte de pierres (*moraine flottante*); une de ces ondulations est visible dans la photographie de la fig. 3 prise



Fig. 3

par moi-même en 1955. Des formations ondulées en sens longitudinal sont aussi visibles pour une des moraines flottantes que je photographiai en 1941, m'étant placée sur le rocher à gauche du glacier, elle est reproduite ici dans la fig. 4; après la glace découverte, située en premier plan, on voit plus loin la moraine flottante qui présente des ondulations le long de l'axe du glacier; ses ondulations sont plus courtes que celles représentées dans les autres photographies, parce qu'elles correspondent à une position différente c'est-à-dire plus éloignée, par rapport au pied de la chute de glace.



Fig. 4

Il faut se rappeler à cet égard que l'ondulation longitudinale de la langue, doit se répéter également dans chaque saison estivale (sans être, toutefois, stationnaire au cours de l'année). Cela est en effet contrôlable dans les différentes photographies que je pris au plateau du Glacier du Lys, au cours de mes visites estivales, désormais nombreuses, et toutes faites au mois d'Août de 1941 jusqu'à 1955. Les vagues près de la chute étaient toujours plus longues que celles qui en étaient plus éloignées. Même sur les nappes de glace découverte des ondulations semblables à celles de moraine flottante peuvent être remarquées, mais elles sont moins accentuées.

Les faits signalés ici peuvent donc être justement définis «vagues-ogives», c'est-à-dire produites par la chute de glace située au-dessus du plateau du glacier. Malheureusement, si le rétrécissement du glacier devait se poursuivre au cours des prochaines années, il ne serait pas possible d'accomplir beaucoup d'études sur ce phénomène si intéressant, à cause du raccourcissement excessif de la langue; toutefois, un accroissement apparent du front pourra être observé lorsque celui-ci se sera déplacé plus en haut, parvenant en correspondance d'une crête d'une des vagues actuellement existantes.

Il est à observer que les nappes de glace découverte, telles qu'on les voit en été sont plus basses que les moraines flottantes puisque chez celles-ci, en raison de leur couverture de détrit, la glace cachée est moins soumise à l'influence de l'ablation, de sorte que le plateau se présente ondulé dans la direction transversale, avec des alternances de veines blanches affaissées et de veines noires surélevées. Ces dernières sont celles qui présentent des ondulations marquées dans le sens longitudinal.

Naples, le 24 Avril 1960.

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INFLUENCE DE LA DYNAMIQUE DES FLEUVES DE GLACE SUR CELLE DE L'INDLANDSIS DU GROENLAND

A. BAUER (EGIG)

RÉSUMÉ

La dynamique de l'Indlandsis du Groenland est influencée par celle des fleuves de glace, par suite de leur mouvement par bloc et par leur décharge sous forme d'icebergs. Le temps de parcours d'une particule de neige jusqu'à la mer est estimé à environ 2000 ans.

SUMMARY

The dynamic of the Greenland ice sheet is influenced by the movement of the ice-streams, because their bloc-movement and ice discharge is larger than ice melting. The travel time of snow until the sea is estimated to be about 2000 years.

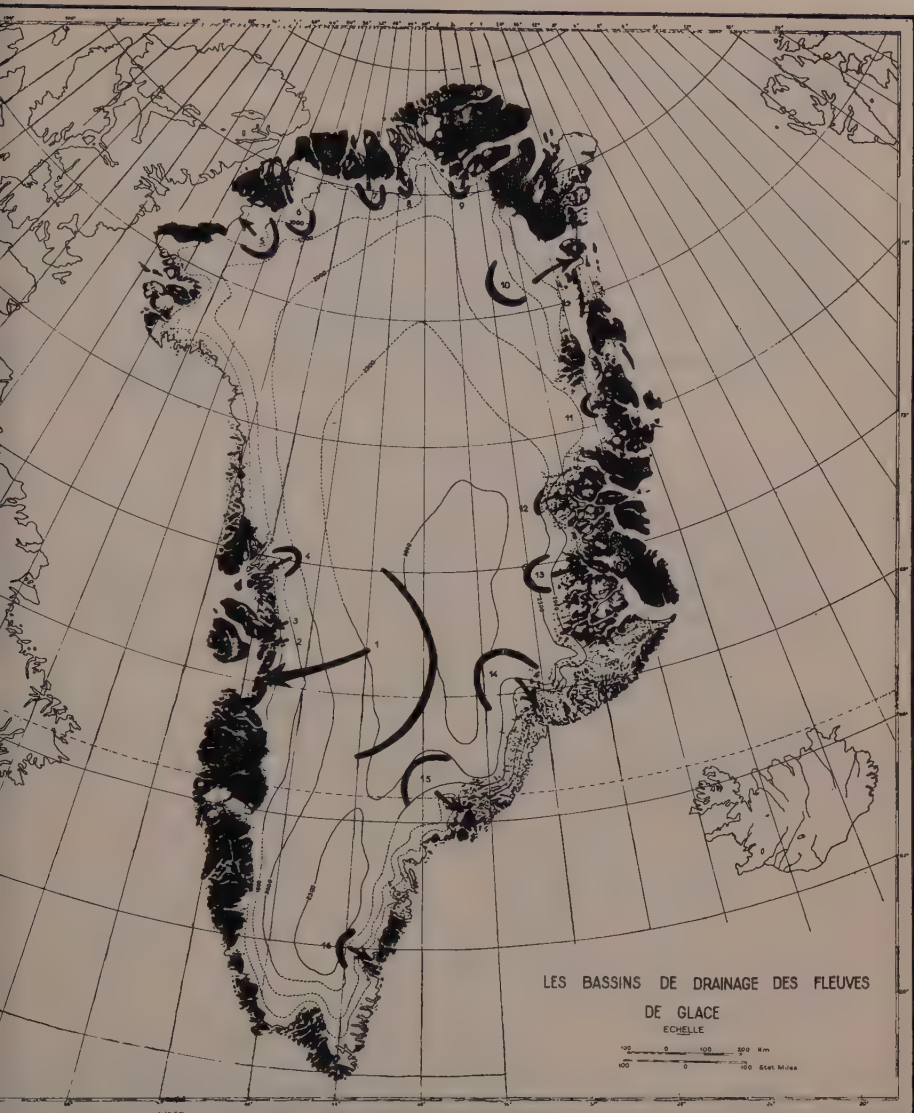
1. INTRODUCTION

Dans notre publication des mensurations du Glacier de l'EQE (Eqip sermia), nous avons pris comme hypothèse de travail la conception de Rink de l'Indlandsis Groenlandais (Bauer, 1955). D'après Rink (Rink, 1857; Drygalski, 1897) les glaces de l'Indlandsis du Groenland sont drainées par une trentaine de glaciers, véritables courants de glace coulant dans la glace plus calme de l'Indlandsis, et appelés fleuves de glace par analogie avec les fleuves des continents. Ces fleuves de glace sont caractérisés par une dépression très crevassée de la surface de l'Indlandsis atteignant plus d'une centaine de kilomètres de longueur jusqu'à la ligne d'équilibre, et au-delà par des zones à grandes crevasses (jusqu'à 100 m d'ouverture) dans le névé jusqu'à 160 km du bord.

L'étude du Glacier de l'EQE nous a montré que ce glacier constituait un tel fleuve de glace. Son mouvement est celui défini par Finsterwalder comme mouvement en bloc (Finsterwalder, 1950, 1951) : Block-Schollen Bewegung. La vitesse du glacier augmente depuis son origine — mal définie — jusqu'à son front, contrairement à la théorie de la dynamique des glaciers et aux observations dans les Alpes. Ces deux particularités s'appliquent à tous les fleuves de glace du Groenland. Dans notre publication des résultats glaciologiques des mesures des Expéditions Polaires Françaises au Groenland, (Bauer, 1954), nous avons montré que la décharge des fleuves de glace du Groenland était de l'ordre de 215 km^3 d'eau par an, presque du même ordre de grandeur que la perte par ablation soit 315 km^3 d'eau par an, pour une accumulation totale de 446 km^3 d'eau par an. En conclusion de cette publication, nous avons attiré l'attention sur l'importance de la perte de substance par décharge sous forme d'icebergs des fleuves de glace sur la dynamique de l'Indlandsis, en essayant de définir le bassin de drainage du plus grand des fleuves de glace : le Jakobshavns Isbrae (fig. 1).

En dehors de la zone immense d'accumulation ($1\,439\,800 \text{ km}^2$ — 83,5 % de la surface totale de l'Indlandsis du Groenland), l'indlandsis du Groenland est caractérisé par deux régions bien distinctes :

- d'une part les zones bordières faiblement crevassées en contact avec leurs moraines frontales où la perte de substance est uniquement due à l'ablation,
- d'autre part les fleuves de glace, véritables courants de glaces dans la glace



Les bassins de drainage des fleuves de glace. (D'après la carte au 1/5.000.000 du Geodactisk Institut, Kobenhavn).

- Glaciers : No.
- 1 Jakobshavn
 - 2 Torssukatak
 - 3 Karajak
 - 4 Rink & Umiamako
 - 5 Humboldt
 - 6 Petermann
 - 7 Ryder
 - 8 Ostfeld
 - 9 Academy
 - 10 Zachariaes & Nioghalvfjerd
 - 11 Heinkel
 - 12 Hamberg
 - 13 Dugaard Jensen
 - 14 Kangerdlugssuaq
 - 15 Sermilik
 - 16 Tingmiarmut

de l'Indlandsis, où la perte de substance, en plus de l'ablation, est due à la décharge sous forme d'icebergs dans les grands fjords, avec les caractéristiques de mouvement déjà mentionnées.

Il faut ajouter que si la vitesse du front calme de l'Indlandsis est de l'ordre de 1 m par an, celle des fronts des fleuves de glace peut atteindre 30 m par jour soit 5475 m par an comme pour le Jakobshavns Isbrae. Les fleuves de glace, par leur décharge sous forme d'icebergs, conditionnent la dynamique de l'Indlandsis Groenlandais. Mouvement par bloc (Koch & Wegener, 1930), accélération du mouvement vers le front des fleuves de glace, perte énorme de substance par décharge sous forme d'icebergs, voilà les caractéristiques réelles qui réduisent à néant les essais modernes de fixer la dynamique des indlandsis d'après la loi de déformation de la glace de Nye (Nye, 1959; Glen, 1955; Haefeli, 1960; Philberth, 1956).

Pour préciser l'influence de la dynamique des fleuves de glace sur celle de l'indlandsis du Groenland, nous allons étudier en particulier le Jakobshavns Isbrae.

2. DETERMINATION DU BASSIN DE DRAINAGE DU JAKOBSHAVNS ISBRAE

Les données dont nous disposons sont les suivantes (Bauer, 1958) :

accumulation moyenne 0,40 m valeur en eau

ablation moyenne 1,10 m valeur en eau

Altitude moyenne de la ligne d'équilibre : 1500 m à 90 km du front décharge sous forme d'icebergs : 20 km³ par an — valeur en eau

(largeur : 7 km

épaisseur moyenne : 0,6 km

vitesse moyenne : 15 m par jour)

Pour avoir une première idée des dimensions du bassin de drainage, nous allons calculer sa longueur en supposant que le glacier est constitué par une bande de 40 km de large. Pour le glacier stationnaire, le bilan annuel donne : (fig. 2).

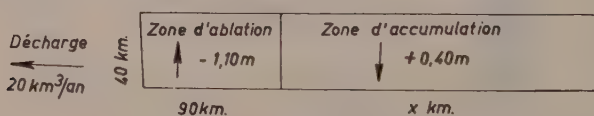


Fig. 2 — Jakobshavns Isbrae

$$20 = 40.90(-1,10.10^{-3}) + 40 \times (0,40.10^{-3})$$

D'où $x = 1500$ km

La longueur totale du glacier dans cette hypothèse serait de l'ordre de 1600 km, c'est-à-dire que l'extrémité de son bassin de drainage devrait être en Islande. Malgré l'in vraisemblance de ce résultat, ce calcul nous donne quand même une idée des dimensions du bassin de drainage.

Reprenons ce calcul avec une hypothèse plus vraisemblable. Nous supposons que, à 10 km du front (figure 3), le bassin de drainage est formé par un secteur circulaire droit. Le ligne d'équilibre est constituée par le rayon $R = 90$ km. Calculons le rayon extrême avec les mêmes données que précédemment (figure 4) :

$$20 = \frac{\pi}{4} (90^2 - 10^2) (-1,10)10^{-3} + \frac{\pi}{4} [(x + 90)^2 - 90^2] (0,40)10^{-3}$$

On en déduit $x = 216$ km, soit une longueur totale du glacier de 306 km.

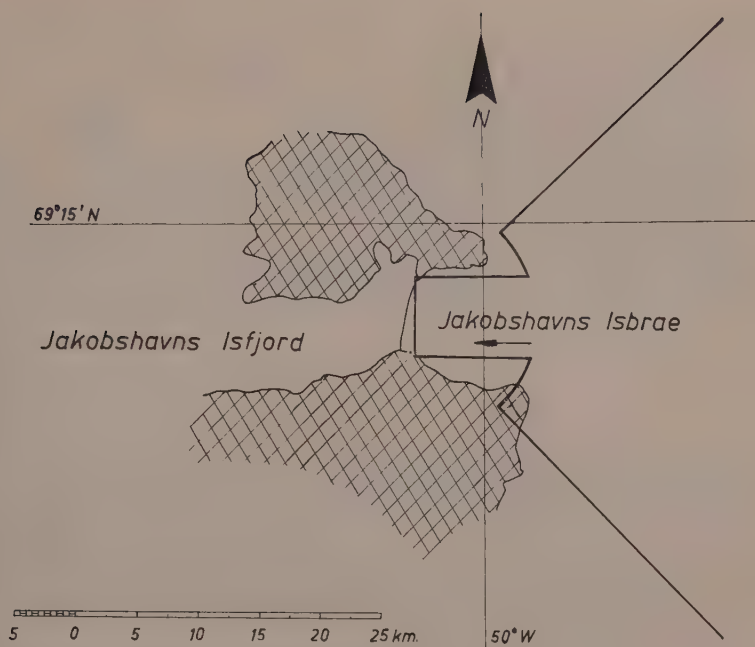


Fig. 3 — Jakobshavns Isbrae

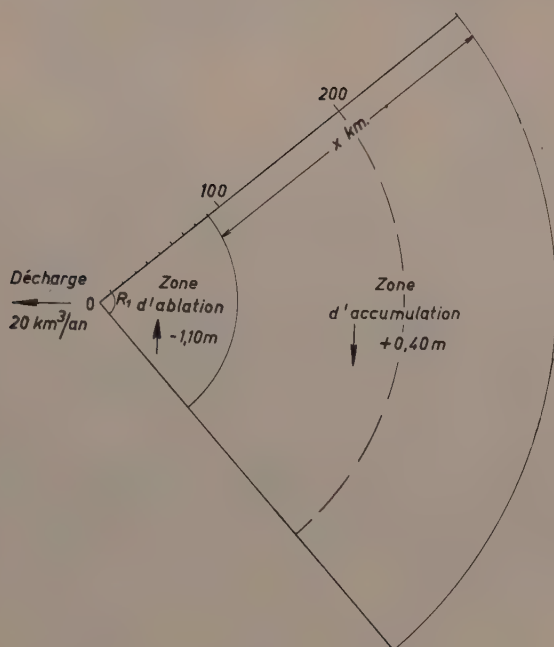
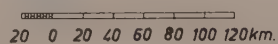


Fig. 4 — Jakobshavns Isbrae — Bassin de drainage.

Vérification :

Surface d'accumulation	67146 km ³	accumulation	26,9 km ³
Surface d'ablation	6283 km ²	ablation	6,9 km ³
Décharge sous forme d'icebergs			20,0 km ³

Nous voyons que pour le Jakobshavns Isbrae, la perte sous forme d'icebergs est trois fois plus grande que la perte de substance par ablation.

L'influence du drainage se fait sentir jusque fort loin dans l'Indlandsis. Ce fait avait déjà été signalé (Bauer, 1954).

De plus, des vallées sous-glaciaires existent sous l'Indlandsis jusqu'à ces distances (Holtzscherer, 1954). Des crevasses peuvent donc exister jusque dans ces régions (Schoumsky, 1955).

3. TEMPS DE PARCOURS D'UNE PARTICULE DE NEIGE JUSQU'À LA MER

Nous pouvons maintenant calculer les vitesses moyennes de tranches du glacier à n'importe quel endroit du bassin de drainage, connaissant l'épaisseur moyenne sur le profil considéré (Holtzscherer, 1954) et supposant le glacier stationnaire.

Nous ne considérons que les vitesses moyennes sur un profil, car cette vitesse moyenne a une réalité pour le transport de la glace, quelle que soit la répartition de la vitesse avec la profondeur. De plus, comme il s'agit d'un mouvement par bloc, cette vitesse moyenne est, au moins dans la dernière partie du glacier, la vitesse réelle de transport (Koch & Wegener, 1930).

Par exemple, sur le profil $R = 90$ km (ligne d'équilibre), il doit passer en un an l'accumulation totale, soit 26,9 km³. La longueur du profil $R = 90$ km est de 141 km. La surface de ce profil, pour une épaisseur moyenne de 1600 m, est de 226 km². La vitesse moyenne est donc de $26,9 : 226 = 0,118$ km par an.

Nous obtenons donc les vitesses moyennes suivantes :

Front $R = 0$ km	v_m 5,475 km/an
$R = 10$ km	1,273
$R = 40$ km	0,280
$R = 90$ km	0,118
$R = 200$ km	0,024
$R = 306$ km	0,000

La courbe des vitesses moyennes nous permet de déterminer la vitesse moyenne générale, soit 0,170 km par an (figure 5). Une particule de neige, tombée à l'extrémité du bassin de drainage mettra donc un temps de l'ordre de 1800 ans pour parvenir à la mer sous forme d'iceberg.

Il est évident qu'il ne s'agit ici que d'un ordre de grandeur, mais nous sommes loin des valeurs invraisemblables de 30.000 à 100.000 ans calculées par Haefeli (1960) et Philberth (1956). Il est à signaler que Hess (1904) était arrivé aux mêmes résultats que nous.

Nos résultats sont en accord avec les résultats de datage de la glace d'iceberg par Scholander (Dansgaard, 1959) à l'aide de C¹⁴ du CO₂ extrait de l'air inclué dans la glace. Ces mesures ont été effectuées au cours de l'Expédition de l'Arctic Institute de D. NUTT en 1958. Les échantillons de glace fondue d'icebergs étaient de 10 tonnes chacun. L'âge de la glace était de l'ordre de 1000 ans, avec un maximum de 3000 ans pour le Glacier d'Upernavik.

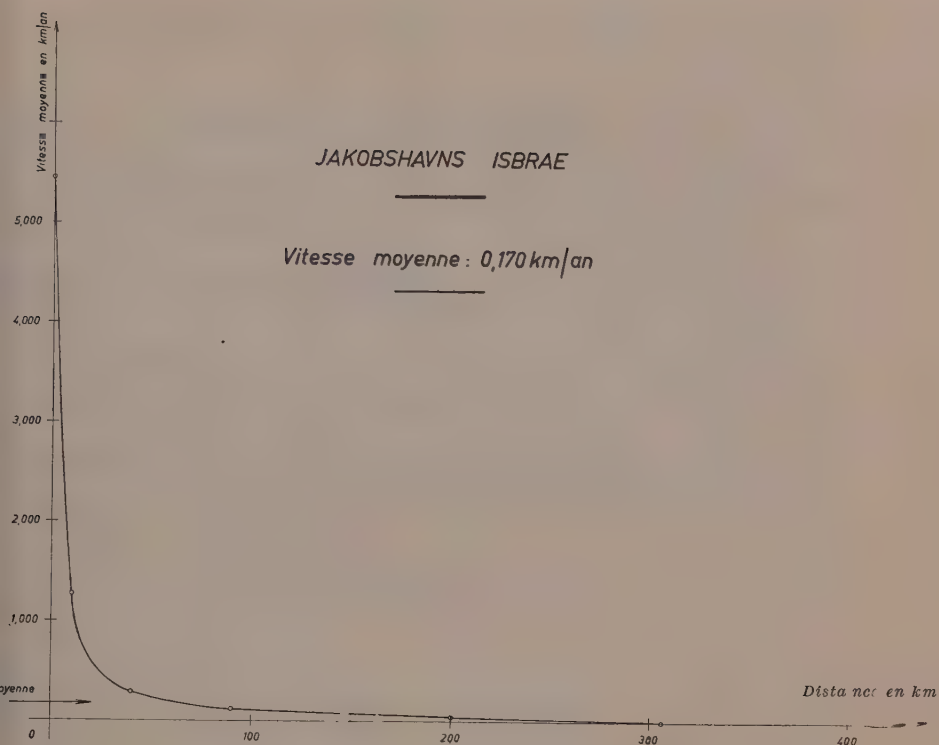


Fig. 5 — Jakobshavn Isbrae — Vitesse moyenne.

4. CONCLUSION

D'une façon générale, nous pouvons conclure que la loi de déformation de la glace de Nye ne s'applique pas aux fleuves de glace, ni aux indlandsis, dont la dynamique est déterminée par la décharge des fleuves de glace qui coulent en eux et drainent leurs glaces jusqu'à la mer.

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